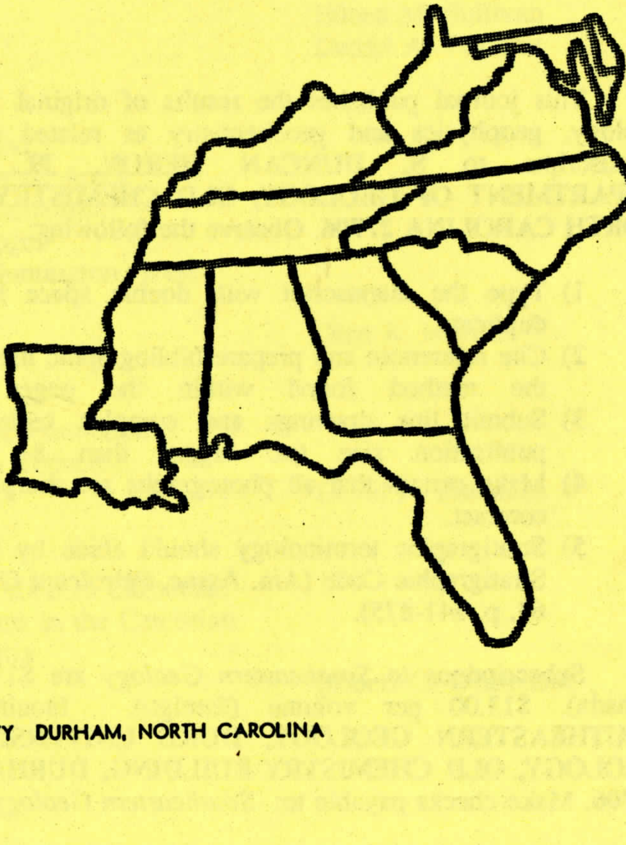
Southeastern Geology: Volume 28, No. 3
February 1988

Edited by: S. Duncan Heron, Jr.

Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

SOUTHEASTERN GEOLOGY



PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 28, NO. 3 FEBRUARY 1988

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

Editor in Chief:
S. Duncan Heron, Jr.

Managing Editor:
James W. Clarke

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SOUTHEASTERN GEOLOGY

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MICROFACIES AND PALEOENVIRONMENTS OF THE MISSISSIPPIAN DENMAR FORMATION, EASTERN WEST VIRGINIA

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ABSTRACT

The Denmar Formation of Randolph County is a package of interbedded marine carbonates and siliciclastics that was deposited on a broad subsiding carbonate platform associated with the syndepositional and intraplatform Beverly uplift. Seven microfacies are recognized: siltstone, dolomitized micrite, pelmicrite, biopelmicrite, quartzose pelmicrite, oosparite and oomicrite. These microfacies characterize three major paleoenvironments: supratidal, intertidal and shallow subtidal.

The Denmar Formation represents a sequence of alternating low energy shallowing-upward pelmicrite-mudstone and higher energy shallowing-upward oosparite-grainstone shelf cycles. The pelmicrite-mudstone cycles are mainly composed of intertidal and supratidal microfacies. The oosparite-grainstone cycles consist of low and higher energy subtidal microfacies.

Three major diagenetic realms are recognized: an early eogenetic, an intermediate eogenetic and a later mesogenetic realm. Diagenesis occurred in a low supratidal flat containing replacement dolomite; a meteoric zone containing blocky calcite spar; a shallow marine phreatic zone with marine cement and micritized fossil fragments; a mixing zone with gypsum nodules and dolomite; and a burial zone containing well-developed stylolites.

INTRODUCTION

Much of the Upper Mississippian Greenbrier Group of the Appalachian basin is located in eastern West Virginia and is exposed in the Appalachian Plateau region. We studied an east-west transect in northern Randolph County, West Virginia where outcrops of the Greenbrier Group exist in several NNE-SSW trending strike belts (Figure 1).

We present and fit petrographic details, including paleoenvironmental and diagenetic interpretations of the lowest formation of the group (Denmar Formation) into the broader stratigraphic framework as reviewed below.

Greenbrier sediments were deposited on a broad carbonate platform in a subsiding basin, and across the adjacent contemporaneously stable shelf (Thomas, 1977) (Figure 2). The lower Greenbrier Denmar Formation, which represents sediment deposition in an overall period of transgression, is composed of interlayered beds of mixed marine carbonate and siliciclastic rocks. Periodic invasion of siliciclastics was also caused by minor reactivation of terrigenous sources to the north, east and southeast of east-central West Virginia. Mixing of siliciclastics was caused by uplift and erosion of previous sediments within the basin along the 38th Parallel Lineament zone, locally termed the Beverly uplift (Yeilding, 1984; Yeilding and others, 1984; Yeilding and Dennison, 1986).

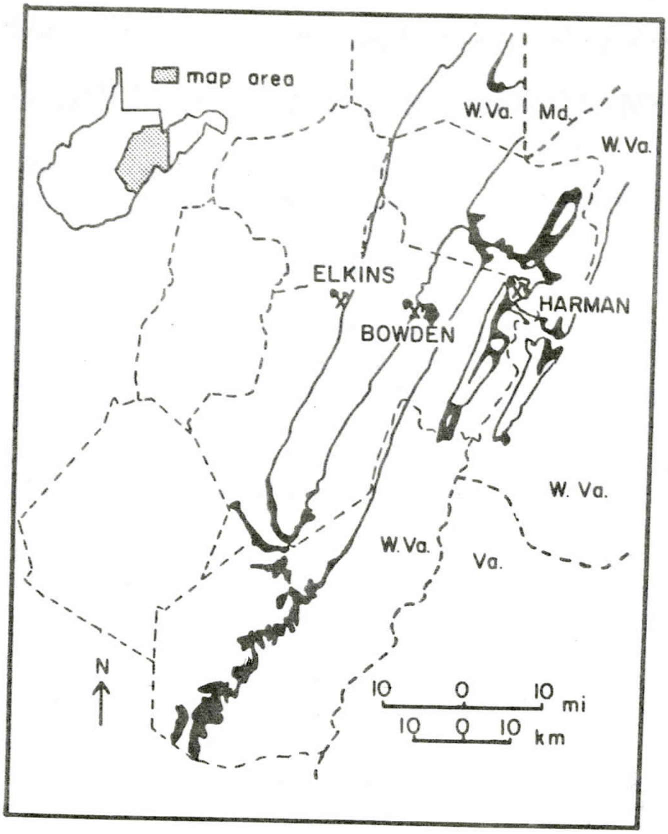


Figure 1. Outcrop of Greenbrier Group in West Virginia (after McCue, Lucke and Woodward, 1939) and location of the three quarries sampled in Randolph County.

The transgressive period represented by the Denmar Formation was terminated by a rapid drop in sea level manifested by the Taggard Formation, a widespread, subaerially exposed and oxidized siltstone bed which caps the Denmar.

METHODS

Three quarry exposures of the Denmar were measured and sampled for petrographic analysis: the Elkins Limestone Company quarry west of Elkins, the Douglas Coal Company quarry at Bowden and the roadside quarry at Harman. Information on the base of the section at Elkins Limestone Quarry was from a well log.

Fifty three 2" x 3" petrographic slides were point-counted using 200 points each, giving a precision of +/– 3.5% for mineral grain abundances in the 50% range (Folk, 1980). Rock textural terms used are from Folk (1962) and Dunham (1962). A random group of samples were selected for X-ray analysis to confirm mineral content. Slides were stained with Alizarin Red S for calcite-dolomite differentiation and with Potassium Ferricyanide for Fe^{+2}. Cathodoluminescent observations were done on a few thin sections, but revealed little.

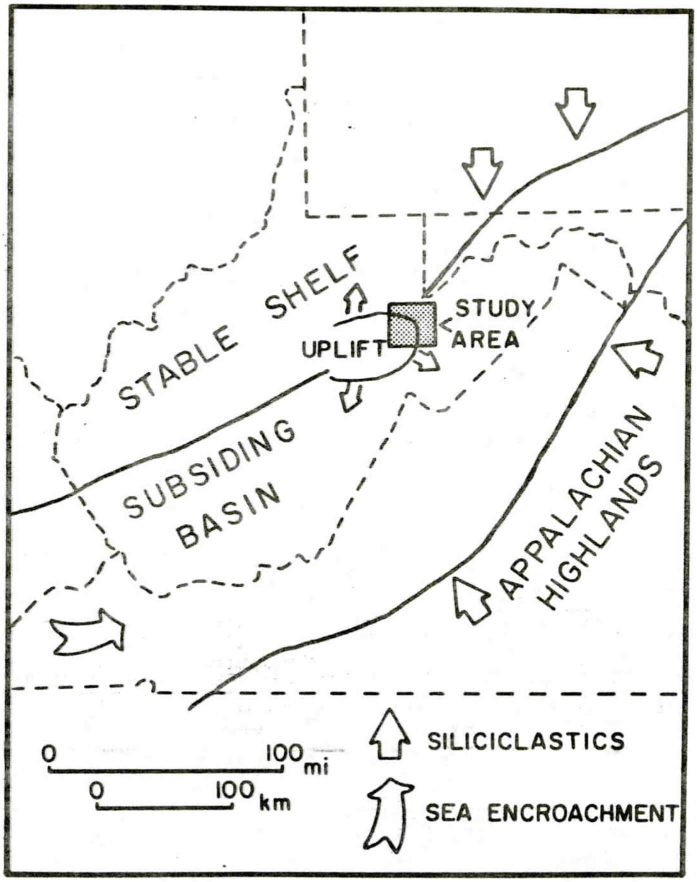


Figure 2. Late Mississippian paleogeography in West Virginia (modified from Donaldson, 1974 and Yeilding, 1984).

STRATIGRAPHIC REVIEW

The Upper Mississippian (Meramecian and lower Chesterian) Greenbrier Group is exposed in Randolph County, West Virginia in a series of NNE-SSW trending strike belts. The Greenbrier Group thickens notably to the south, and thins to the north-central part of Randolph County. This thinning coincides, and apparently is associated, with the 38th Parallel Lineament zone of Dennison and Dever (1976), Werner (1976) and Yeilding and Dennison (1986).

The Denmar Formation was originally not recognized in the region (Wells, 1950). Recent stratigraphic studies by Yeilding (1984), however, reveal that it is present. In Randolph County, the Denmar unconformably overlies the Upper Devonian Hampshire Formation, and the Lower Mississippian Pocono and Maccrady Formations (Yeilding, 1984). Overlying the Denmar is the terrigenous Taggard Formation and the predominantly carbonate Pickaway, Union and Alderson Formations. Together, these comprise the Greenbrier Group (Figure 3).

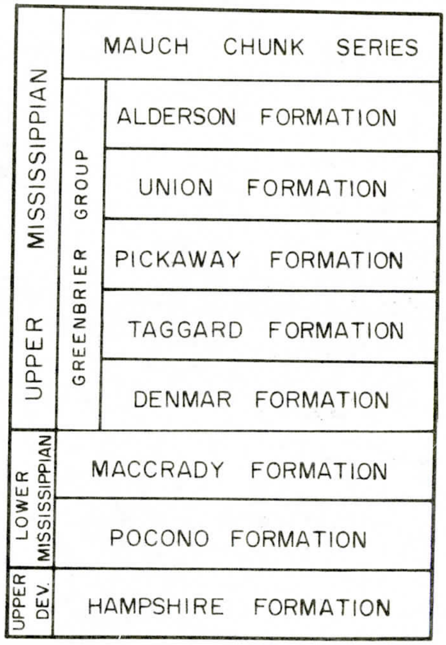


Figure 3. Devonian and Mississippian strata, northern Randolph County, West Virginia. Vertical scale does not represent thickness.

MICROFACIES

Seven microfacies are recognized: siltstone, dolomitized micrite, pelmicrite, biopelmicrite, quartzose pelmicrite, oosparite and oomicrite. Early and intermediate diagenesis are subdivisions of the eogenetic realm; burial diagenesis represents the mesogenetic realm. The diagenetic realms are from Choquette and Pray (1970).

Siltstone

This microfacies is a dolomitic, calcitic quartz siltstone, ranging in color from grayish red to dark gray (Figure 4a). Samples display moderate sorting, have a low allochemical content (<13%), and contain a small to moderate amount of intergranular carbonate mud and cement. The allochems consist of a small number of fossil fragments, peloids consisting of fecal pellets and micritized foraminifera, rare rip-up clasts or ooids, and blebs of hydrocarbons. These allochems are often poorly preserved because of silt interpenetration, dolomitic replacement and compaction.

Early diagenesis: Micrite recrystallized to microspar and some replacement of micrite and quartz by dolomite.

Intermediate diagenesis: Inter-granular voids filled with single meteoric spar crystals.

Burial diagenesis: Compaction caused slight interpenetration of silt and allochems.

Dolomitized Micrite

Dolomite, in this hydrocarbon-rich microfacies, occurs as euhedral to subhedral rhombs which vary from .007-.15 mm (Figure 4b). These rocks range in color from light gray to dark gray, and usually have a low allochemical content (<1%), although in some rocks ghost foraminifera and pellets can be discerned as well as a few ostracod and crinoid fragments. Samples commonly have some quartz silt, a few rip-up clasts, a very small amount of intra-allochemical cement, blebs of hydrocarbons, planar fenestrae, anhedral pyrite and burrow swirls.

Early diagenesis: Extensive penecontemporaneous dolomitization of the micrite and some allochems in the supratidal zone. Meteoric calcite spar filled fenestrae.

Intermediate diagenesis: Period of mixing zone dolomitization developed larger euhedral to anhedral crystals which replaced some calcite spar (Figure 4c).

Burial diagenesis: Stylolites and fractures formed. Fractures filled with meteoric calcite spar.

Pelmicrite

This microfacies is a hydrocarbon-bearing, partially dolomitized pelmicrite (Figure 4d). These rocks range in color from medium light dark to dark gray. Pellets and micritized foraminifera comprise peloids which vary from .025 to .1 mm. Most samples contain more than 5% fossil allochems, which include ostracod, trilobite, crinoid, bryozoan, brachiopod, gastropod and pelecypod fragments. Some are surrounded by micrite envelopes. Other allochems are intraclasts, superficial ooids and true ooids. Many samples contain some quartz silt, blebs of hydrocarbons and a small amount of pyrite. Some samples contain geodes or fenestrae filled with calcite spar. Euhedral to anhedral dolomite replaces some peloidal micrite. There is usually less than 5% inter- and intra-allochemical calcite spar.

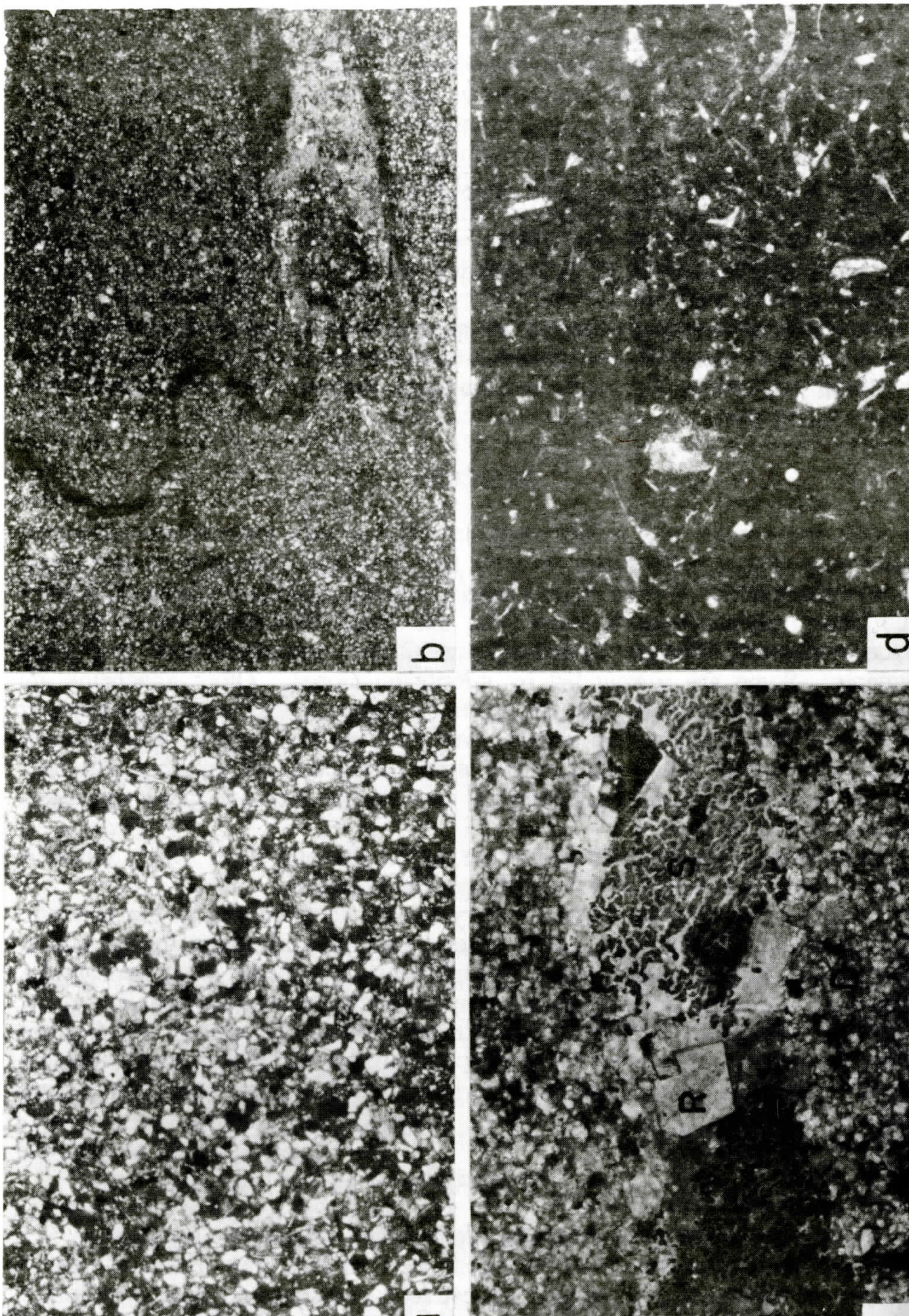
Early diagenesis: Micritization in the marine phreatic zone. Many bioclasts developed micrite envelopes or were completely micritized.

Intermediate diagenesis: Micrite recrystallized to microspar in the meteoric zone. Equant calcite spar filled inter- and intra-allochem pore spaces. Hydrocarbons migrated into voids terminating spar growth in some samples (Figure 5a). Anhedral dolomite replaced some calcite micrite and spar.

Burial diagenesis: Stylolites and fractures formed. Dissolution and fracture porosity developed. Fractures filled with equant spar or "ropey" calcite cement (Figure 5b), which may represent a period of rock movement which caused cement deformation.

Biopelmicrite

This microfacies is a partially dolomitized biopelmicrite (Figure 5c). Samples range in color from medium gray to medium dark gray. Fossil allochems are 11% of the rock and consist of ostracod, brachiopod, crinoid, trilobite, bryozoan, foraminifera, pelecypod and alga, along with intraclasts composed of peloids and fossil fragments, peloids, superficial ooids and true ooids. Peloids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are fecal pellets and micritized foraminifera. All samples contain some inter- and intra-allochemical cement, while some have calcite spar-filled geodes after gypsum. Burrows are commonly present.

Early diagenesis: Micritization in the marine phreatic zone formed micrite envelopes on many bioclasts, and complete micritization of foraminifera tests formed peloids. Evaporitic gypsum nodules developed in the mixing zone followed by dolomite replacement of pelmicrite matrix (Figure 5d). Dolomite replaced some micrite and quartz grains.

Intermediate diagenesis: Calcite replaced some gypsum nodules (Figure 6a), while in others gypsum was leached and equant calcite spar was precipitated into the void. Subhedral to anhedral pyrite formed.

Burial diagenesis: Some compaction and stylolite formation occurred.

Quartzose Pelmicrite

This microfacies is a slightly dolomitized quartzose pelmicrite (Figure 6b). It ranges in color from very light gray to dark gray, commonly intertonguing. In outcrop, the base of the unit appears as thin discontinuous subparallel light and dark layers. Darker beds are caused by a higher concentration of hydrocarbons. Higher in the section, the unit becomes mottled to the point of no longer being bedded. Peloids are mainly micritized foraminifera, which are mixed with a subequal amount of quartz sand. Intergranular voids are filled with microspar.

Early diagenesis: Total micritization of foraminifera formed peloids in the marine phreatic zone. Dolomite replaced some quartz (Figure 6c).

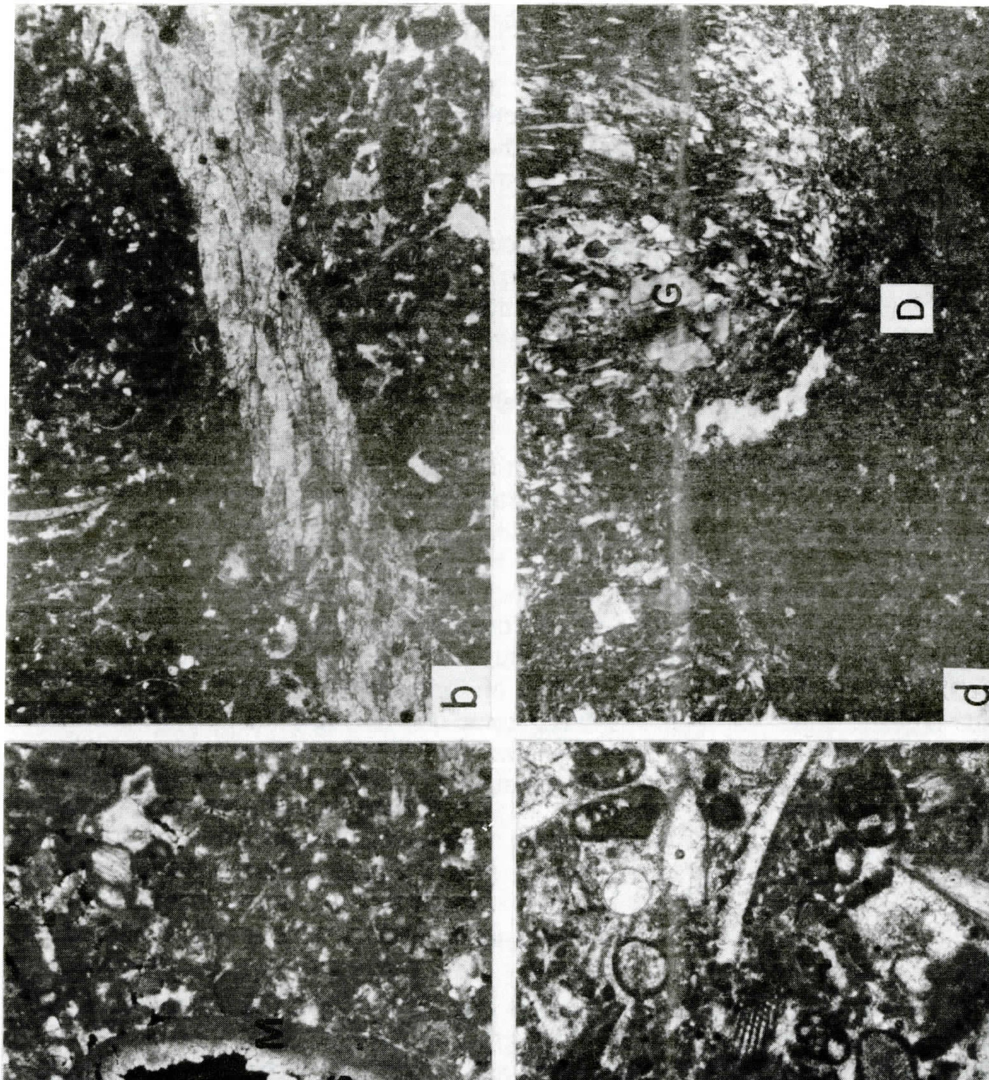
Intermediate diagenesis: Inter-granular micrite recrystallized to microspar. Small inter-allochem voids filled with equant calcite spar. Some peloidal micrite was replaced by dolomite.

Burial diagenesis: Fractures formed, which were later filled by hydrocarbons.

Oosparite

This partially dolomitized microfacies is cross-bedded in outcrop, and ranges in color from grayish red purple to dark gray. Most ooids are .1 to .5 mm and contain crinoid, foraminifera, ostracod, pelecypod, algal, bryozoan, brachiopod,

Figure 4. a) Terrigenous siltstone. Silt is terrigenous quartz. Also included are peloids, micrite, clay and hydrocarbons. Dolomite replaced some terrigenous and carbonate grains. Photograph is 2.7 mm wide. b) Dolomitized micrite. Dolomite rhombs replaced carbonate mud. Calcite spar filled fracture. Stylolite concentrated hydrocarbons. Possible ghost foraminifera in matrix. Photograph is 4 mm wide. c) Euhedral dolomite rhombs replaced calcite in spar-filled fracture in dolomitized micrite. Calcite spar within fracture is stained with Alizarin Red S (medium gray in photograph). Dark areas are hydrocarbon-rich. S=spar, D=dolomite matrix, R=dolomite rhomb. Photograph is 1 mm wide, crossed nicols. d) Pelmicrite. Peloids are fecal pellets and micritized foraminifera. Microfacies also includes fossil fragments (ostracod, pelecypod, crinoid), quartz silt and hydrocarbons. Dolomite replaced some carbonate and quartz grains. Photograph is 4 mm wide.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trilobite and gastropod fragments as nuclei (Figure 6d). Both superficial and true ooids are found in most samples mixed with intraclasts composed of ooids, peloids and shell fragments, peloids of pellets and micritized foraminifera and some fossil fragments, many with micrite envelopes. The ooid coatings are predominantly tangential, and some are inter-mixed with radial coats. Allochems are moderately well preserved. Most samples show inter-layering of carbonate allochems and quartz sand and silt, and all are cemented by inter- and intra-allochemical cement.

Early diagenesis: Micritization and micrite envelope development of most free bioclasts occurred in the marine phreatic zone. Some foraminifera were completely micritized forming peloids (Figure 7a). Intra-allochem isopachous rims of fibrous marine cement grew in a few ostracod valves. Syntaxial overgrowths formed around some crinoid fragments.

Intermediate diagenesis: Micrite recrystallized to microspar. Aragonite bioclasts (including ooid nuclei) were leached forming moldic porosity. Inter- and intra-allochem pore spaces filled with equant meteoric calcite spar. Subhedral to anhedral dolomite replaced micrite and some spar. Pyrite formed between some ooids and peloids.

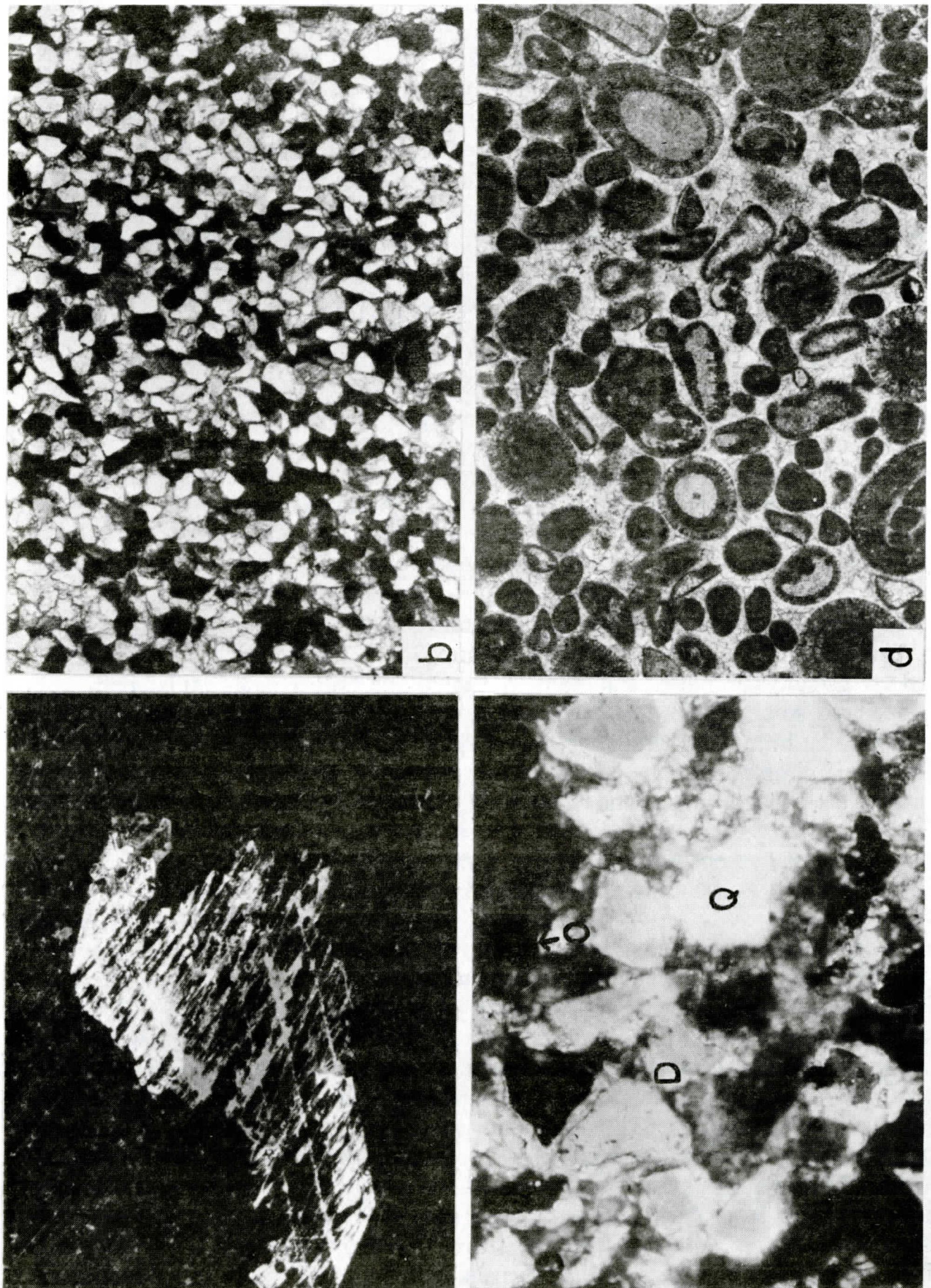
Burial diagenesis: Stylolites and fractures formed. Fracture dissolution and porosity developed and filled with equant or "ropey" calcite spar.

Oomicrite

This microfacies is a partially dolomitized, peloidal oomicrite, ranging in color from medium gray to dark gray (Figure 7b). The rocks are often wavy bedded to pseudo-laminated and contain true ooids, superficial ooids, peloids, intraclasts and some interbedded quartz silt. Micritized foraminifera and pellets form peloids. Intraclasts are agglutinated peloids, ooids and shell fragments. Ooid nuclei consist mainly of crinoid, ostracod, formainifera, pelecypod, brachiopod, bryozoan, algal and gastropod fragments. There is come intra-allochemical spar, but material between the allochems is recrystallized micrite. Some dolomite is present as replacement of micrite.

Early diagenesis: Micrite envelope formation affected most bioclasts in the

Figure 5. a) Totally micritized shell (within intraclast) enveloping internal cavity now filled with calcite spar and hydrocarbon in pelmicrite. H=hydrocarbon, S=spar, M=micritized shell. Photograph is 2 mm wide. b) Calcite fracture-filling in pelmicrite. Ropey spar morphology may have been caused by rock movement during cement formation. Dark round features are air bubbles. Photograph is 4 mm wide. c) Biopelmicrite. Peloids are fecal pellets and micritized foraminifera. Microfacies also includes fossil fragments (ostracod, pelecypod, crinoid, alga, foraminifera, brachiopod, bryozoan), superficial ooids, quartz silt and hydrocarbons. Inter-allochem voids are filled with micrite and calcite spar. Dolomite replaces some carbonate and quartz grains. Photograph is 4 mm wide. d) Gypsum nodule and dolomite rhombs from Denmar Formation biopelmicrite south of field area (Fairfax Sand and Crushed Stone Quarry near Beverly, West Virginia). Gypsum appears to have formed by "displacive" growth. G=gypsum, D=dolomitized biopelmicrite. Photograph is 4 mm wide, crossed nicols.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marine phreatic zone. Total micritization of some foraminifera formed peloids. Syntaxial overgrowths formed on some crinoid fragments. Dolomite replaced some micrite and quartz silt.

Intermediate diagenesis: Micrite recrystallized to microspar. Inter- and intra-allochem voids filled with equant meteoric calcite spar. Euhedral to anhedral dolomite replaced some inter-allochem micrite. Pyrite formed “sooty” films on some ooids and peloids.

Burial diagenesis: Stylolites formed.

PALEOENVIRONMENTS

With the use of outcrop features, petrography, stratigraphic sequence and recent analogs we interpret that three major juxtaposed paleoenvironments existed in east-central West Virginia during Late Mississippian Denmar sedimentation (Figure 8). These are supratidal, intertidal, and shallow subtidal environments discussed below.

Supratidal Environment

The supratidal environment is typically an extensive flat area situated just above normal high tide, adjacent to a marine zone. This marine zone can directly effect the supratidal sediment during spring tides or storms. Common features include micrite, mudcracks, planar fenestrae or caliche crusts, algal laminations, rip-up clasts and siliciclastics (Shinn, 1983). Shelled organisms may be rare, but occasional burrows are seen (Textoris, 1968). Supratidal sediments also may have a reddish color caused by oxidation. In arid climates, evaporite minerals such as gypsum, anhydrite, celestite or halite may form (Shinn, 1983; Wilson, 1975), or brines will form and affect lower beds.

Eight samples, which include the dolomitized micrite and siltstone micro-facies, have been assigned to the supratidal environment. The dolomite is typically euhedral, less than .015 mm and formed early. These samples contain rare (<1%) fossil allochems (usually ostracods), planar fenestrae and rip-up

Figure 6. a) Remnant gypsum crystal(s) (white) partially replaced by calcite in biopelmicrite. Calcite spar is stained with Alizarin Red S (dark gray). Photograph is 4 mm wide. b) Quartzose pelmicrite. Peloids are mainly micritized foraminifera and some fecal pellets. Terrigenous material is quartz sand and silt. Sediment also includes a small number of fossil fragments and hydrocarbon blebs. Dolomite replaced many quartz grains and some carbonate peloids. A small amount of microspar filled inter-granular voids. Photograph is 4 mm wide. c) Dolomite replaced quartz grains in quartzose pelmicrite. Inter-granular calcite is stained with Alizarin Red S (medium gray). Dolomite forms corrosive rims around quartz. Peloids are micritized foraminifera. Dark gray material is hydrocarbons. Q=quartz, D=dolomite, O=hydrocarbon material. Photograph is 2 mm wide, crossed nicols. d) Oosparite, which includes tangential and radial ooids, superficial ooids, peloids mainly composed of micritized foraminifera, fossil fragments (crinoid, foraminifera), intraclasts and some quartz silt. Inter-allochem voids are filled with calcite spar. Photograph is 4 mm wide.

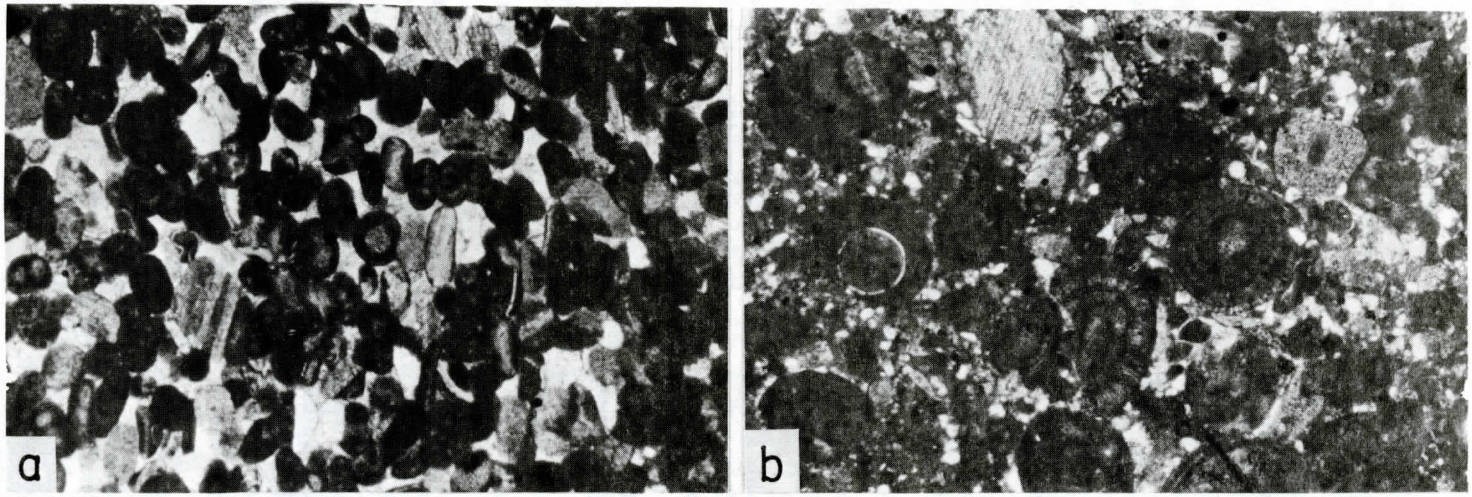

Figure 7. a) Micritized foraminifera now as peloids in oosparite. Many peloids still show foraminferal features while others are micritized beyond recognition. Photograph is 4 mm wide. b) Oomicrite. Microfacies includes ooids, superficial ooids, peloids, crinoid fragments, quartz sand and silt and some hydrocarbons. Inter-allochem voids are filled with microspar. Dolomite replaced some quartz and carbonate grains. Photograph is 4 mm wide.

and intraclasts composed of ooids and peloids. Other characteristics remain the same throughout the intertidal zone.

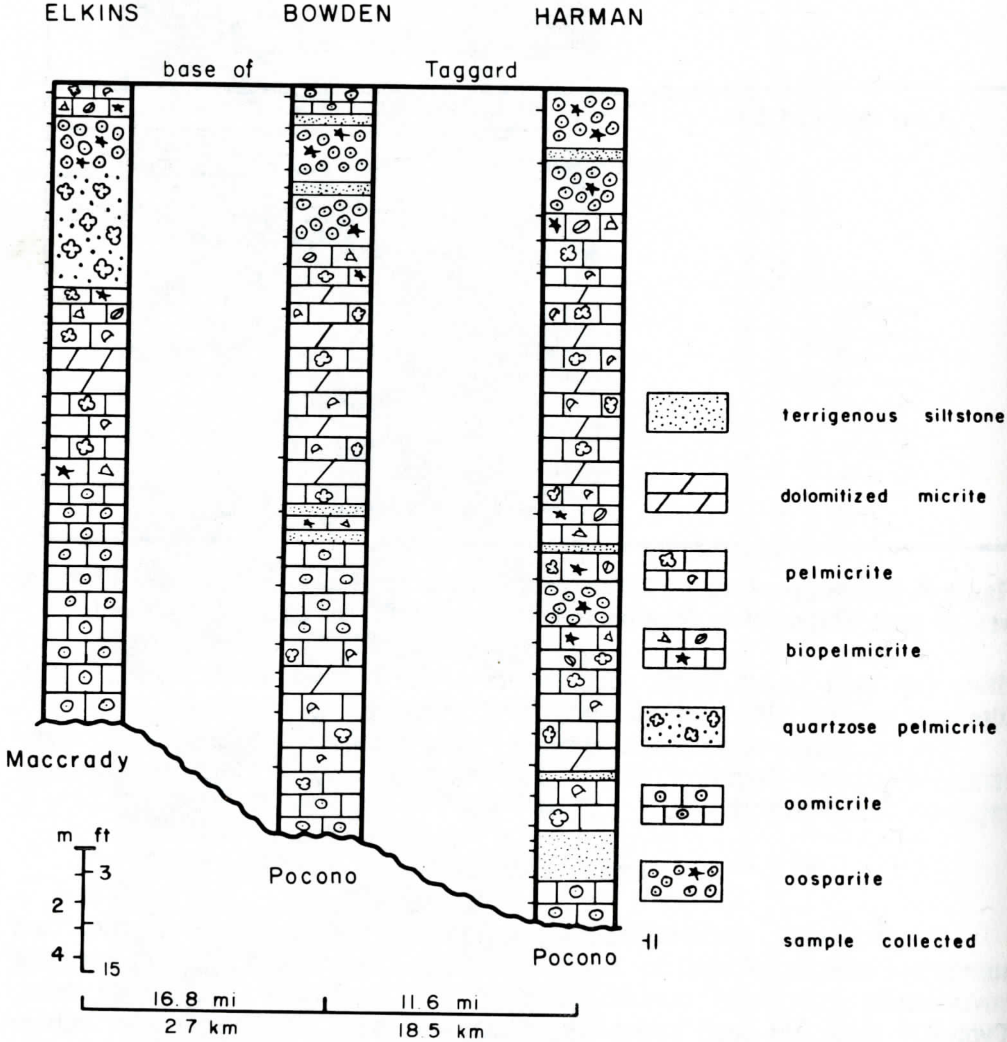

Figure 9. Detailed stratigraphic diagram for quarries sampled, Denmar Formation, northern Randolph County, West Virginia.

Shallow Subtidal Environment

The shallow subtidal environment is juxtaposed to the intertidal environment and may include low to high energy zones. Low energy areas are typically characterized by peloidal mud, diverse and abundant biota, boring algae and fungi, and burrows (Enos, 1983). Higher energy areas exhibit sorted skeletal and peloidal sands with very little micrite. Ooids may form (James, 1977). Cross-bedding is common in higher energy zones (Wilson, 1975).

Twenty-four samples have been assigned to the subtidal environment, 15 in a lower energy zone and nine in a higher energy zone. Microfacies representing

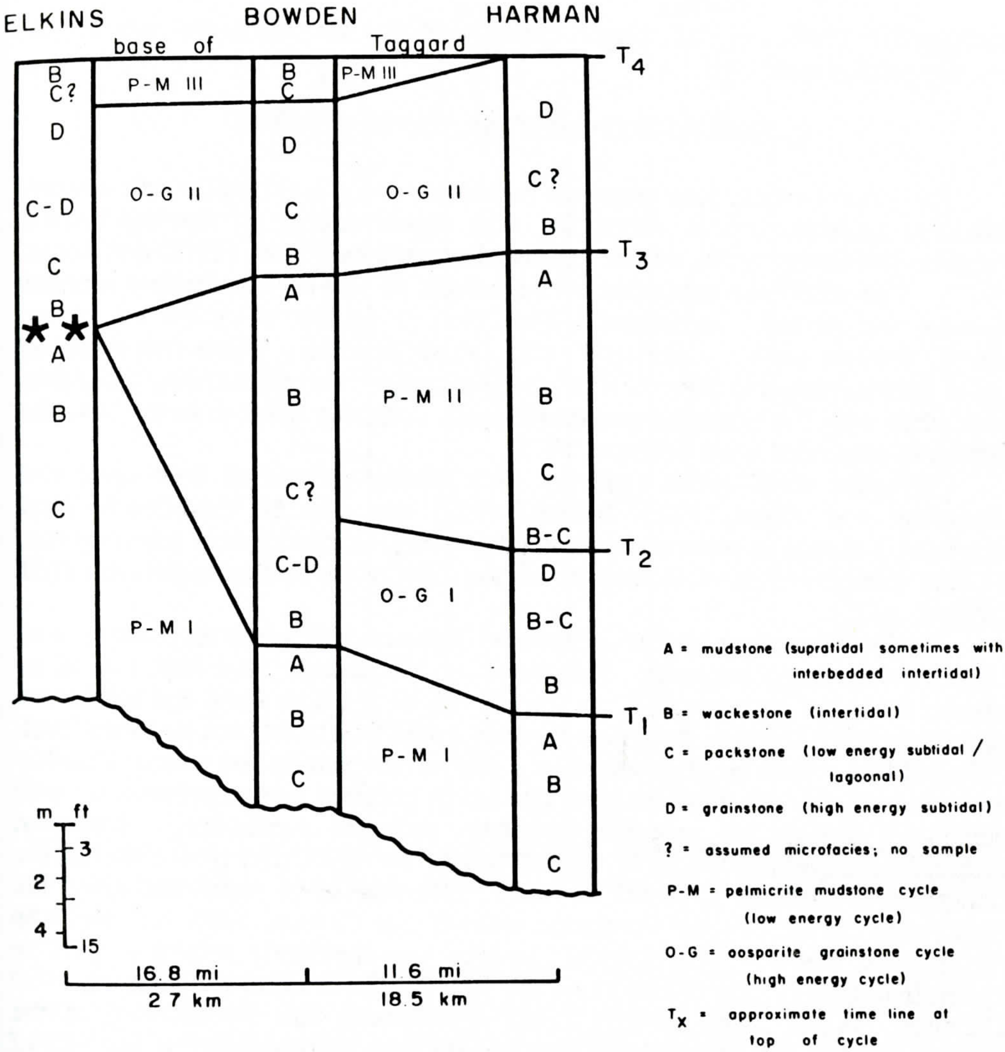

Figure 10. Cycle correlation diagram for Denmar Formation, northern Randolph County, West Virginia. **This probably represents a series of diastems, not a single disconformity, which formed during exposure and activity along the 38th Parallel Lineament.

low energy, possibly lagoonal environments, are biopelmicrites, oomicrites and quartzose pelmicrites. They are characterized by pelleting, burrowing, intraclasts of agglutinated peloids, diverse and abundant bioclasts, and abundant micrite. Many samples have a large amount of siliciclastics mixed with the allochems. Gypsum nodules, or their calcite replacements, are common and suggest exposure and aridity in a juxtaposed environment to provide brines for mixing.

Oosparites represent the high energy shoal portion of the environment. They are cross-bedded in outcrop, and in thin section are characterized by peloids (many of which are micritized foraminifera), superficial ooids, true ooids, intraclasts of agglutinated peloids and ooids and a few bioclast fragments (mainly crinoid and brachiopod). Some samples contain quartz sand and silt which are interbedded with the peloids and ooids.

PALEOENVIRONMENTAL DEVELOPMENT

The entire Denmar rock sequence represents low-energy, shallowing-upward pelmicrite-mudstone cycles alternating with higher energy, shallowing-upward oosparite-grainstone cycles similar to those discussed by Wilson (1975) and James (1977). The pelmicrite-mudstone cycles consist of low-energy shallow subtidal lagoonal and peritidal carbonates. The oosparite-grainstone cycles are higher energy, shallow shelf-lagoonal and oolitic shoal deposits. These two types of cycles are interpreted to form on wide, shallow cratonic shelves and may grade into each other with the oosparite-grainstone cycle occurring more seaward than the pelmicrite-mudstone cycle (Wilson, 1975).

Carbonate shelf cycles typically form during periods of base level rise (Anderson and others, 1978; Bebout, 1977) and can be modified by local tectonism, a change in water circulation, tidal variations, overall shelf submergence, climatic changes, sea level fluctuation and the relief of the carbonate platform itself (Wilson, 1975).

The Denmar cycles appear to represent a sequence of minor regressions in an overall transgressive sequence. This would not necessarily have been caused by eustatic sea level fluctuation. If sea level remained at a high stand and subsidence occurred within the basin, carbonate deposits would then build back up to sea level. This build-up would be represented as a regressive cycle in the rocks. Possibly there was a steady but slow sea level rise which occurred contemporaneously with punctuated periods of repeated carbonate sediment production. Basically, carbonate sediment productivity lagged behind sea level rise; thus allowing the transgressive Denmar sequence to form, followed by a rapid sea level fall represented by the Taggard Formation redbeds (see Carozzi, 1986, for details on this eustatic model). The peritidal complex was apparently related to positive syntectonic activity near Elkins along the 38th Parallel fracture zone (Yeilding and Dennison, 1986). The Beverly uplift is indicated also by thinning of the Greenbrier Group in eastern West Virginia and by a disconformity at the base of the Denmar (Yeilding and Dennison, 1986).

Figure 9 shows the distribution of the seven microfacies for the Denmar Formation of northern Randolph County. Some appear to be traceable throughout the three sections, while others are not. Cycles, which were established on the basis of the microfacies and their associated environments, also appear to be correlative (Figure 10).

The basal cycle in all three sections is a low energy pelmicrite-mudstone cycle (P-M I) capped by a supratidal unit. The easterly Harman and Bowden sections next show an increase in water energy forming an oosparite-grainstone cycle (O-G I). This cycle is followed by another pelmicrite-mudstone cycle (P-M II). The westernmost Elkins section exhibits an anomalous cycle pattern. The initial cycle is a low energy pelmicrite-mudstone unit as the other two sections. But, at time T2, the Elkins area did not shift to the subtidal environment and remained in the supratidal environment.

Yeilding and Dennison (1986) propose that tectonic uplift along the 38th

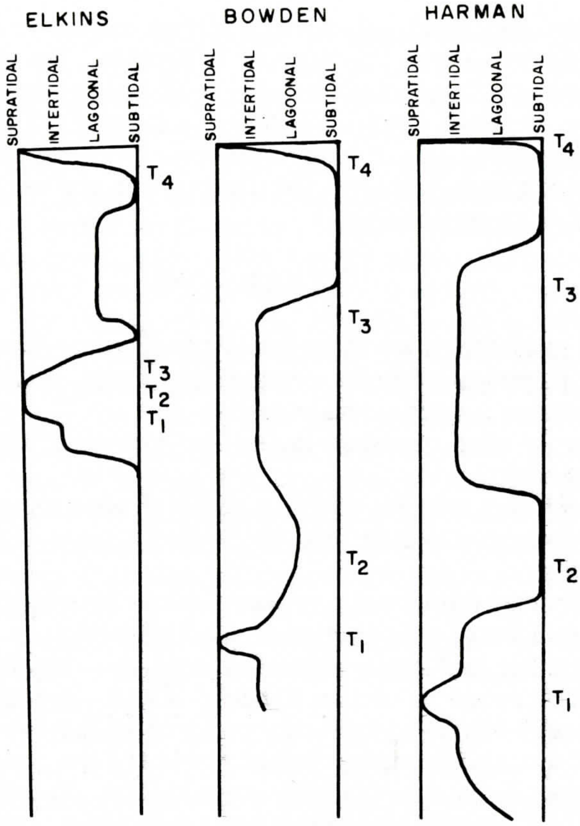

Figure 11. Paleoenvironments associated with transgressive-regressive phases of Denmar Formation. See Figure 10 for placement of time (T) designations within cycles.

Parallel Lineament growth feature (Beverly uplift) in western Randolph County occurred during the time of Denmar sedimentation. The petrographic and cycle correlation diagrams appear to confirm this. Basal carbonates at the Elkins Quarry are subtidal in origin indicating tectonic growth did not occur until after Denmar sedimentation had begun. Once started, tectonic growth continued, causing the development of an initial pelmicrite- mudstone cycle which is capped by approximately three meters of supratidal rocks. After tectonic growth terminated, the Elkins area was flooded and subtidal deposition began again as in the other two areas. All three sections exhibit an high energy oosparite-grainstone cycle at this point (O-G II). The easternmost Harman quarry is stratigraphically terminated by the Taggard redbeds while the other two sections began another pelmicrite-mudstone cycle (P-M III), before being covered by Taggard silts.

Further paleoenvironmental reconstruction (Figure 11) is based on approximate time lines shown in Figure 10. The tops of the cycles in the Bowden and Harman sections are used as time lines because the cycles are successive and include approximately the same thickness of sediment in each section. Because of the tectonic growth period at Elkins during cycles O-G I and P-M II, this westernmost section is treated as supratidal at times T1, T2 and T3. A time line, T4, is drawn where the uppermost pelmicrite-mudstone cycles at Elkins and Bowden appear to be correlative with the Taggard at Harman. For this area of northern Randolph County, the Taggard can be used as an approximate time line, based on cycle correlation (Sullivan, 1985). Actually, the Taggard redbeds could represent a cap (peritidal exposure) on all three sections. This regionally deposited unit represents a penultimate sea level fall (regression) and the end of the overall Denmar carbonate transgression.

CONCLUSIONS

1) Seven microfacies are identified in the Denmar Formation in Randolph County, West Virginia: siltstone, dolomitized micrite, pelmicrite, biopelmicrite, quartzose pelmicrite, oomicrite and oosparite.

2) Sediments were deposited in the low supratidal, intertidal and shallow subtidal environments.

3) The rock sequence represents a series of alternating upward shallowing low energy pelmicrite-mudstone and shallowing upward high energy oosparite-grainstone cycles.

4) The Denmar represents a series of minor regressions within a regional transgression, terminated quickly by a regional regression.

5) Petrographic and cycle correlation appears to support the proposal that tectonic growth occurred in western Randolph County in early Denmar time.

6) Diagenesis of these sediments occurred in the low supratidal flat, the meteoric zone, the shallow marine phreatic zone and the burial zone.

7) Diagenetic changes include micritization, intra-allochemical marine cement, gypsum growth, dolomitization, freshwater equant spar, leaching developing moldic and vuggy porosity, micrite recrystallization, replacement of gypsum and quartz by calcite and dolomite, pyrite growth, compaction, stylolitization, fracturing and hydrocarbon migration.

ACKNOWLEDGEMENTS

We are grateful for the exchange of geologic information and ideas with John M. Dennison, Cindy A. Yeilding, and A. Conrad Neumann. The research was funded by grants from the Appalachian Basin Industrial Associates, the American Association of Petroleum Geologists, the West Virginia Geological and Economic Survey, the University of North Carolina Geology Martin Trust Fund, and the University of North Carolina Research Council.

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MARINE TRANSGRESSION AND SYNDEPOSITIONAL TECTONICS; AMES MEMBER (GLENSHAW FORMATION, CONEMAUGH GROUP, UPPER CARBONIFEROUS) NEAR HUNTINGTON, WEST VIRGINIA

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ABSTRACT

The Ames Member of the Glenshaw Formation, Conemaugh Group (Virgilian, Upper Carboniferous) is the product of the last significant marine invasion in the Paleozoic history of eastern North America. In the area near Huntington, West Virginia, there are not fewer than five, and possibly six, correspondences between geometry of rock bodies within the Ames and the geologic structures in or on which they are located. These thickenings and thinnings within the Ames indicate that the southwestern part of this widespread marine unit was deposited during subsidence resulting from tectonic movement along the axis of a major regional east-west to northeast-southwest structure (the Pittsburgh-Parkersburg-Huntington Syncline). This movement probably admitted and localized marine waters from the north or northwest across the preexisting broad alluvial plain. This downwarping represented a pulse in the progressive tectonism of the Alleghanian Orogeny and a possible reactivation of basement tectonic activity along the 38th Parallel lineament. Exposures in and around Huntington, West Virginia, demonstrate lithogenetic responses to this tectonism within the Ames marine environments, demonstrating that development of this structure had begun by Late Pennsylvanian time.

INTRODUCTION AND GENERAL STRATIGRAPHY

For decades the "Appalachian Revolution", now the Alleghanian Orogeny, was considered to be the ultimate Paleozoic event in eastern North America (e.g. Dunbar, 1949:251, 281). It was interpreted to have been "an act of violence that itself brought on the close of the Paleozoic Era" (quote from Spieker, 1956, p. 1775, who debunked these ideas) and clearly post-Carboniferous in age because of the gentle deformation of Permian (?) strata (Dunkard Group) in the Allegheny Plateau. Woodward (1957) in defining the Alleghanian Orogeny, concluded that much, perhaps most, of the movement had to have occurred during Carboniferous time. Modern radiometric dating techniques (Glover, and others, 1983; Dallmeyer, and others, 1986) along with analysis of sedimentary sources and a better understanding of orogenic events in the setting of plate tectonics, have combined to modify that age assignment. It is now clear that initiation of the Alleghanian Orogeny with its dynamic movement, intrusion, and metamorphism of the eastern margin of the North American Plate began at least as early as the middle of the Carboniferous (Thomas and Mack, 1982) and continued progressively to near the end of the Paleozoic (Secor, and others, 1986). Thus the Permo-Carboniferous rocks represent thick clastic wedges shed from the rising mountains (Ferm, 1974).

Continuing deformation led to increased sediment supply, to progradation of proximal over distal depositional settings, and to increasing mild deformation of the clastic wedges in West Virginia and Pennsylvania (Williams and Bragonier, 1974) that probably progressed northwesterly through time. Luz, and others (1984), assign an age to the Ames of 288 Ma. In view of the currently suggested date for the Pennsylvanian-Permian boundary of 286 Ma (Palmer, 1983) that would make the Ames very young Virgilian. Although not directly comparable, this late Virgilian assignment is compatible with that based on conodont evidence (Merrill, 1974). This general interval of time was one of considerable Alleghanian tectonic activity corresponding in a general way to the Clarks Hill (D3) folding (Secor, and others, 1986).

The repetitive intercalation of marine strata into dominantly non-marine Carboniferous rocks has given rise to numerous speculations about the causitive mechanism(s) for this alternation. In the Illinois Basin and Midcontinent, where a substantial portion of the succession is marine and some marine intervals are geographically widespread, Weller (1956) suggested that minor tectonic movements would account for the numerous marine transgressions and regressions. Alternatively, Wanless and Shepard (1936) suggested that variation in sea level caused by waxing and waning glaciers in the southern hemisphere could produce the same effect. This idea has been most recently amplified by Ross and Ross (1985) and Heckel (1986). In the lower part of the coal measures in the Appalachian region, where strata with marine fossils are generally much less widespread, Ferm (1970), while not excluding sea level control, proposed that differential rates of sedimentation associated with delta building could produce recurrence of strata with marine fossils in the stratigraphic record. Unless episodes of southern hemisphere glaciation/deglaciation can be matched precisely to particular northern hemisphere marine invasions, Ferm concluded that it would appear there is no easy way to resolve this problem. Brown (1969) reached nearly the same conclusions dealing with delta- and carbonate bank-related deposits in north-central Texas.

Relative rise in sea level can also result from sedimentary compaction, the likely cause for most of the Appalachian Pennsylvanian transgressions that produced marine deposits of relatively limited areal extent (Ferm, 1970). For more widespread deposits such as the Ames Member of the Glenshaw Formation, however, a different mechanism must be sought. Evidence that would tie particular glaciations with eustatic events is lacking, and statements that assert synchroneity because of biostratigraphic similarities, are undemonstrated or unconvincing (Ross and Ross, 1985). Refinement of biostratigraphic zonations is simply not sufficiently precise at the present time. Moreover, repeating such assertions of undemonstrated eustasy based upon assumed synchroneity becomes an essentially unprovable hypothesis, and worse, this discourages search for evidence to support the leading alternative hypothesis: tectonism.

An opportunity to document a tectonically triggered transgression in a part of the Ames basin and evaluate results of syndepositional tectonism was offered by recent highway construction in the tri-state region of western West Virginia, southern Ohio, and northeastern Kentucky near Huntington, West Virginia (Figure 1 and Table 1). Much of the general geology and lithogenesis of the strata are summarized in Merrill (1986), but that paper does not consider the syndepositional tectonics implied by the variations within the Ames Member. Strata exposed near Huntington are the Conemaugh Group that is stratigraphically above the major coal-bearing succession in much of the Appalachian region and thus record depositional events and environments that followed those described by Ferm, and others (1971) nearby in eastern Kentucky. The Conemaugh Group near Huntington is composed mostly of non-marine silt-shale, sandstone, and claystone (much of it red), but the lower half (Glenshaw Formation) contains at least four zones yielding marine fossils; in ascending order, Lower Brush Creek, Upper Brush Creek, Cambridge and Ames Members. The Ames is widely distributed within the Appalachian Basin and over most of its outcrop area it generally represents an episode of minimum detrital influx coinciding with the widespread distribution of marine water. In the Huntington area it is dominantly terrigenous, however, and numerous excellent exposures provide an outstanding opportunity for studies of possible effects of structure and detrital influx upon marine sedimentation.

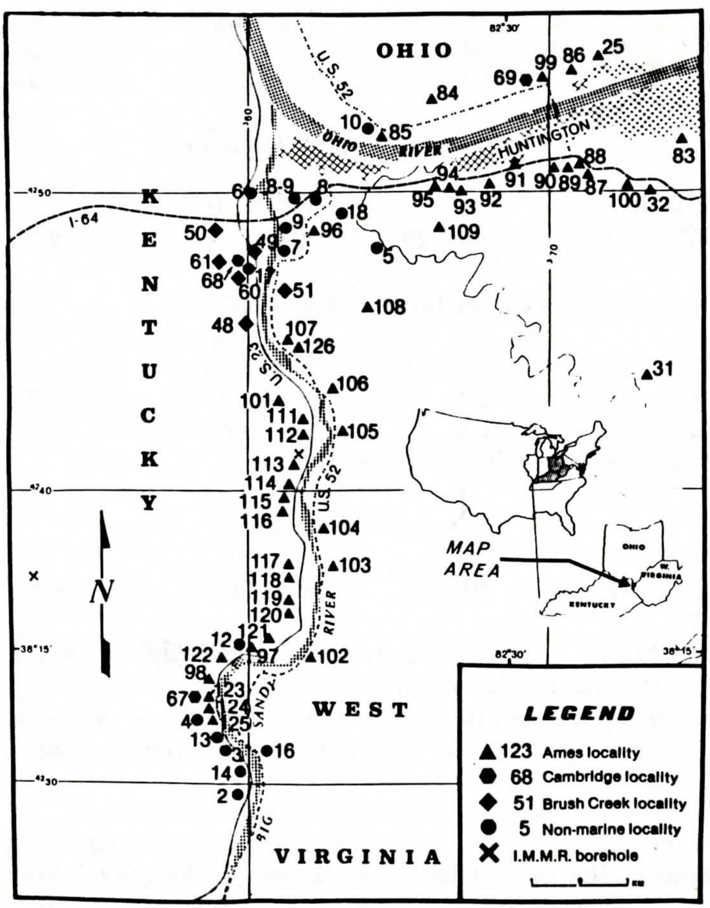

Figure 1. Locality map of measured sections in the Huntington area. Grid is the UTM 10,000 m grid within 100,000 m grid square 17SLN. I.M.M.R. is the Kentucky Institute for Mining and Minerals Research.

Table 1. Supplementary locality register of newer localities not listed in Merrill, 1986 (p. 158) or shown on Figure 1. GSO# = Ohio Division of Geological Survey stratigraphic file number. States are abbreviated per postal abbreviations. Other abbreviations, in order of appearance, counties: Boyd, Lawrence, and Cabell; township-district, Union; quadrangles; Burnaugh, Huntington, Lavalette, Catlettsburg; roads, U.S. 23, Ohio 7.

Loc	*GSO#*	*ST*	*CO*	*TP/DIS*	*QU*	*RD*	*UTMCoord (17SLN)*	*N lat° ' "*	*W long° ' "*
				Ames Localities					
141	16697	KY	BO	------	BU	23	59174595	38 21 09	82 36 42
142	16698	KY	BO	------	BU	23	59374556	38 20 56	82 36 33
143	16699	KY	BO	------	BU	23	59984502	38 21 40	82 36 07
144	16700	KY	BO	------	BU	23	60244466	38 20 28	82 35 57
145	16701	KY	BO	------	BU	23	60534426	38 20 16	82 35 44
146	16703	KY	BO	------	BU	23	61754280	38 19 28	82 34 52
147	16778	OH	LA	UN	HU	7	72265433	38 25 49	82 27 48
148	16779	OH	LA	UN	HU	---	72965489	38 26 06	82 27 17
150	16780	WV	CA	UN	LA	---	74734367	38 20 04	82 26 00
				Upper Brush Creek Localities					
60	16694	KY	BO	------	BU	23	59514782	38 22 09	82 36 31
61	16695	KY	BO	------	BU	23	59347721	38 21 49	82 36 36
				Lower Brush Creek Localities					
67	16692	KY	BO	------	CA	23	59704944	38 22 31	82 36 23
68	16693	KY	BO	------	BU	23	59684929	38 22 24	82 36 24
69	16696	KY	BO	------	BU	23	59154668	38 21 24	82 36 43
70	Same locality as Upper Brush Creek 61								
				Non-marine Locality					
19	16702	KY	BO	------	BU	23	61034358	38 19 53	82 35 23

Merrill (1986, Figure 1, p. 158) listed three Cambridge localities. Conodont evidence indicates that locality 67 is Cambridge, but 68 is Upper Brush Creek. Insufficient data do not permit confirmation or rejection of the correlation of locality 69. "Cambridge" locality 68 has been redesignated Upper Brush Creek locality 62.

Regional Geology of the Ames Member and Regional Structure

The Ames Member is the stratigraphically highest persistent marine bed in the Pennsylvanian, indeed the Paleozoic, of the Appalachian Basin. It is found from its type section in southeastern Ohio across the eastern Ohio outcrop into

Pennsylvania as far east as the Allegheny Front, then southward into western Maryland and across northern West Virgina. It is not present along the Conemaugh outcrop in central West Virginia, but reappears in the Huntington area of northwestern West Virginia and northeastern Kentucky. The general trend of the southern shoreline of the Ames marine incursion, as it is known from surface and subsurface information, is shown on Figure 2. Through much of its northwestern outcrop and subcrop in eastern Ohio and western Pennsylvania, the Ames is represented by a prominent limestone on the order of 0.3-1.7 m (one to five ft) in thickness overlain by a few centimetres of fossiliferous shale. Southward through Ohio and West Virginia, limestone is less abundant, but it is replaced by shale with marine fossils. From central West Virginia westward to Louisa, Kentucky, marine, or even brackish water, fossils are absent at the position of the Ames Member and it seems reasonable to assume that this absence of fossils reflects the Ames shoreline having been located north of the present Conemaugh outcrop belt. Ames marine fossils are found in westernmost West Virginia and in the northern part of eastern Kentucky, but they are not present south of the line shown on Figure 2.

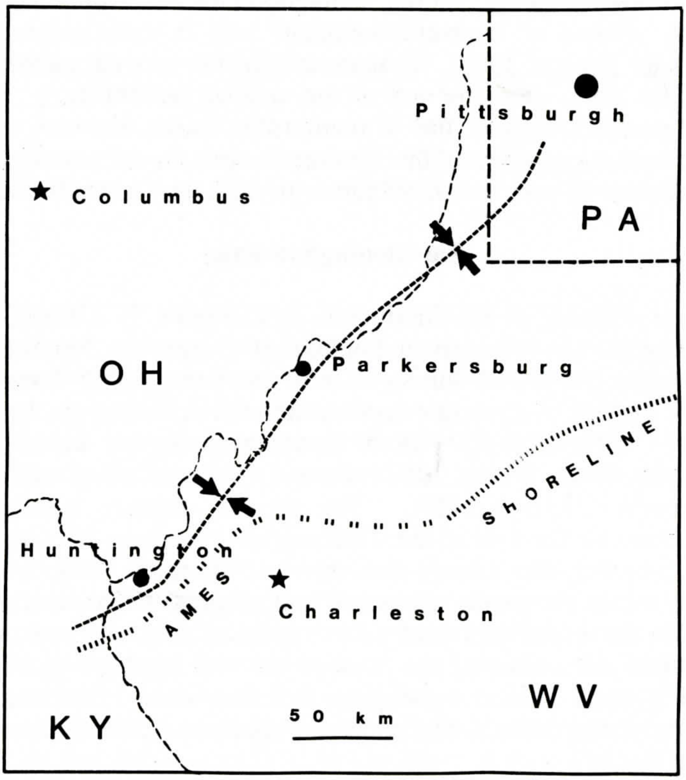

Figure 2. Approximate location of the southern Ames shoreline in western West Virginia in relation to axis of the Pittsburgh-Parkersburg-Huntington Syncline. Wider spacing of symbols reflects poorer data in the subsurface. Shoreline data based on Grimsley, 1910; Hennen, 1911, 1912, 1917; Hennen and Reger, 1914; Krebs, 1911; Krebs and Teets, 1913, 1914, 1915; Merrill, 1986; Reger, 1916, Reger and Teets, 1918; and Strubble, and others, 1971.

In the immediate vicinity of Huntington, the area underlain by the Ames is astride the axis of the Pittsburgh-Parkersburg-Huntington Syncline. This axis coincides with the axis of the Dunkard Basin and the center of the entire Allegheny Plateau. On the northwest flank of this fold from northwestern Pennsylvania through southern Ohio the dips are gentle ($<1/3°$)to the southeast and east and are probably the result of regional subsidence. On the eastern outcrop in southwestern Pennsylvania and northern West Virginia the Ames is found in a number of gentle folds that involve strata as young as the Dunkard Group (Permian?). These folds parallel late Paleozoic folds in central Pennsylvania, western Maryland and eastern West Virginia (Williams and Bragonier, 1974). On the southern limb of the syncline in the study area, which is the approximate position of the Ames shoreline, the structure has a north-northwestward dip with a strike that is parallel to structures crossing southern West Virginia and eastern Kentucky. These parallel structures farther south in eastern Kentucky include a series of thrust faults. Studies of older coal-bearing rocks in eastern Kentucky and southern West Virginia have shown that these rocks represent a great wedge of northwestward-prograding deltaic sediments (Ferm and Horne, 1979). These are cut by a series of east-northeastward trending faults contemporaneous with deposition that controlled the volume of deposited sediment and in some cases directions of sediment transfer (Horne, 1979). Concurrence of this general pattern with that of the trend of the Ames shoreline and of the axis of the Pittsburg- Parkersburgh-Huntington Syncline suggests that a relationship exists between them and that details of Ames sedimentation in the Huntington area should provide a further test for tectonic control of sediment distribution in the Late Carboniferous.

The Huntington Area

Structural mapping of the Huntington area (Figure 3) is based on data from exposures along the newly constructed highways (Figure 1). Previous mapping by the West Virginia Geological Survey (Krebs and Teets, 1913; Cross, and others, 1956) showed the axis of the major regional structure crossing the Big Sandy River at about UTM 17SLN626375. More recent mapping by the U.S. Geological Survey (Spenser, 1964; Sharps, 1967) showed the same axis crossing the river 5 km further north (17SLN623429). The present mapping confirms the more northerly location of the axis of the Pittsburgh-Parkersburg-Huntington Syncline. The present mapping also shows that dips on the north flank of the structure approaching the axis are greatly steepened in contrast to the regional strucuture ($<3\ 1/2°$). Within the broad and nearly flat structural low, however, there are two distinct synclinal axes crossing the river at the two locations previously shown, separated by a small northeast-plunging anticlinal nose. The southern limb is steeper than the northern limb, with measured apparent dips of as much as $7°$ to the north (Figure 3).

Structural Relationships of Stratigraphic Units

There are at least five and perhaps six significant relationships between the geometries of Ames lithosomes and structural elements in the Huntington area. 1) The maximum thickness of the Ames marine unit coincides with the axis of the

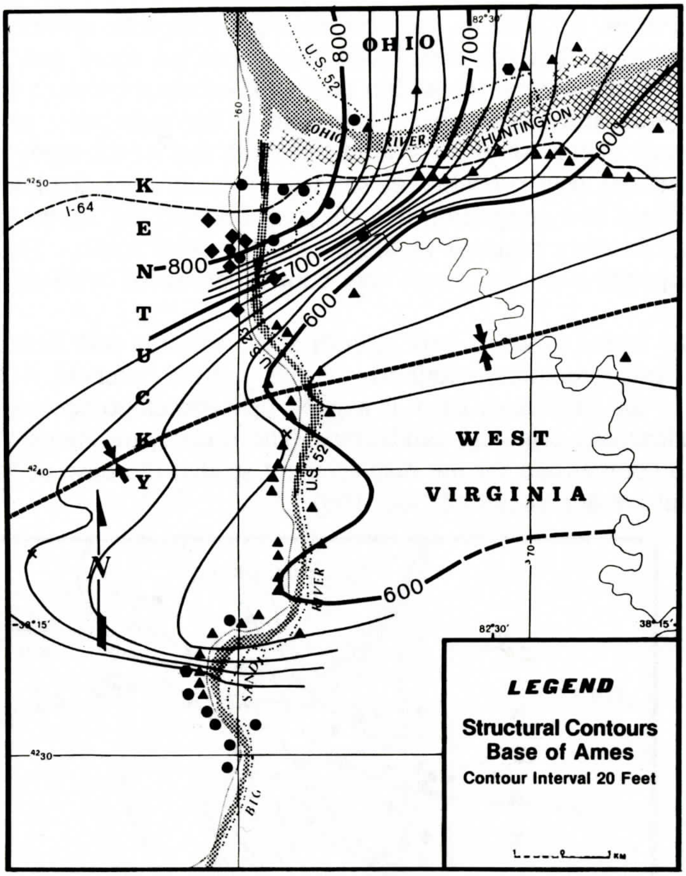

Figure 3. Structural contour map of the base of the Ames Member in the Huntington area. Elevations in feet above mean sea level permitting comparison of the structural contours with the topographic contours on existing U.S. Geological Survey topographic maps. Syncline axis shown is that of the Pittsburg-Parkersburg-Huntington Syncline. Compare with Figure 1 for identification of points.

Pittsburgh-Parkersburg-Huntington Syncline; 2) the maximum thickness of calcareous crinoidal sandstone also occurs along the primary syncline axis; 3) the calcareous crinoidal sandstone pinches out on the anticlinal nose south of this axis, the thickness dropping from its maximum of nearly 5.5 m (18 ft) to zero in less than two kilometres (1.1 mi); 4) the total Ames marine interval thins from a maximum of more than 9.0 m (30 ft) to less than a third that (just over 2 m, about 8 ft) on the crest of this gentle anticline; 5) the Ames marine interval thickens once again into the secondary syncline (located where the West Virginia Geological Survey maps show the Pittsburgh-Parkersburg-Huntington Syncline) regaining a thickness of more than 3.0 m (12 ft); 6) the Ames interval appears to be thinning once again updip on the steeply dipping southern limb of the syncline. These relationships are shown by comparing Figures 3 and 5 as well as Figure 6. Marine rocks at the Ames position do not extend as far south as the Louisa quadrangle

(Connor and Flores, 1978). In the absence of a traceable erosion surface, the logical conclusion is that the Ames was not deposited there and the southern shoreline was located someplace within the ten kilometres between the south end of this study area and the Louisa quadrangle. The extremely steep dips and thinning tendency of the Ames interval at the south end of the study area indicate that updip pinchout was probably closer to the study area than to the Louisa quadrangle. There is a suggestion that the coal beneath the Ames shows similar thickening and thinning patterns to those of the marine rocks. Hence it would appear that deposition of the Ames was contemporaneous with development of these structures.

Along the major syncline axis in both West Virginia and Kentucky there is evidence of flow structures within the Ames marine interval. On the West Virginia side of the Big Sandy there is a small penecontemporaneous reverse fault cutting the calcareous crinoidal sandstone. The underlying Ames shales exhibit plastic flow to compensate for the displacement so that the base of the Ames and underlying coal are not displaced (loc. 106).

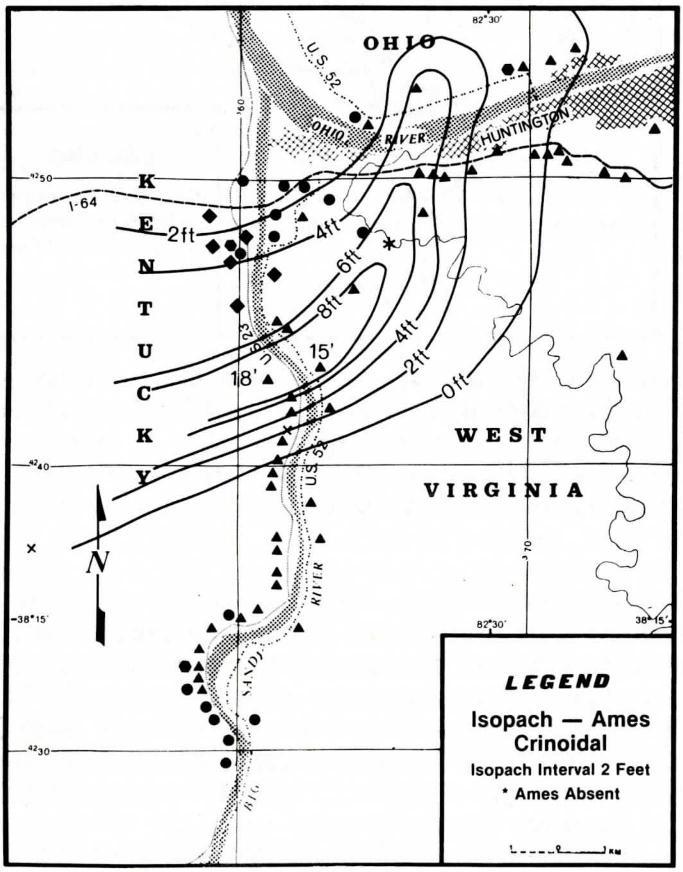

Figure 4. Isopach map of the calcareous crinoidal sandstone lithosome of the Ames Member in the Huntington area. Local absence of Ames Member assumed to be the result of postdepsotional erosion and not contoured as zero. Compare with Figure 1 and Figure 3.

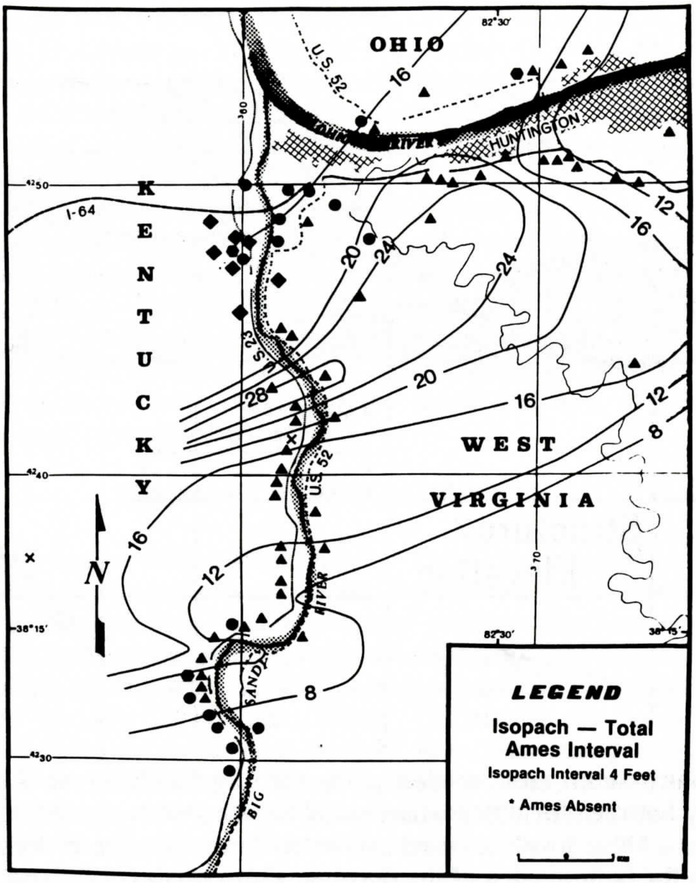

Figure 5. Isopach map of the total thickness of the Ames Member in the Huntington area. Comments as for Figure 4.

The correspondences between structures and rock bodies indicate that downwarping was contemporaneous with Ames deposition near Huntington. If it can be assumed that the eastward dip in southern Ohio represents regional tilting subsequent to Ames deposition, and that the steepened dip encountered as this north flank approaches the axis of the syncline dates from the time of deposition, then it can be speculated that the broad flat eastward plunging bottom of the syncline itself may be a composite structure. Its steep flanks would have been formed, in part at least, at the time of deposition, but with the plunge inherited from subsequent eastward tilting. The steepness of the flanks suggests that the syncline may be an east-northeast trending graben-like structure, related to structures further to the south.

There is substantial evidence for Carboniferous tectonism in the area. Horne (1979) and Dever (1977) summarized the known contemporaneous structure during Carboniferous deposition in eastern Kentucky. Prior to deltaic progradation across northeastern Kentucky, earlier Carboniferous carbonate environments were made shallower or emergent by gentle upward movements along the Waverly Arch. During later Carboniferous time a series of down-to-the-

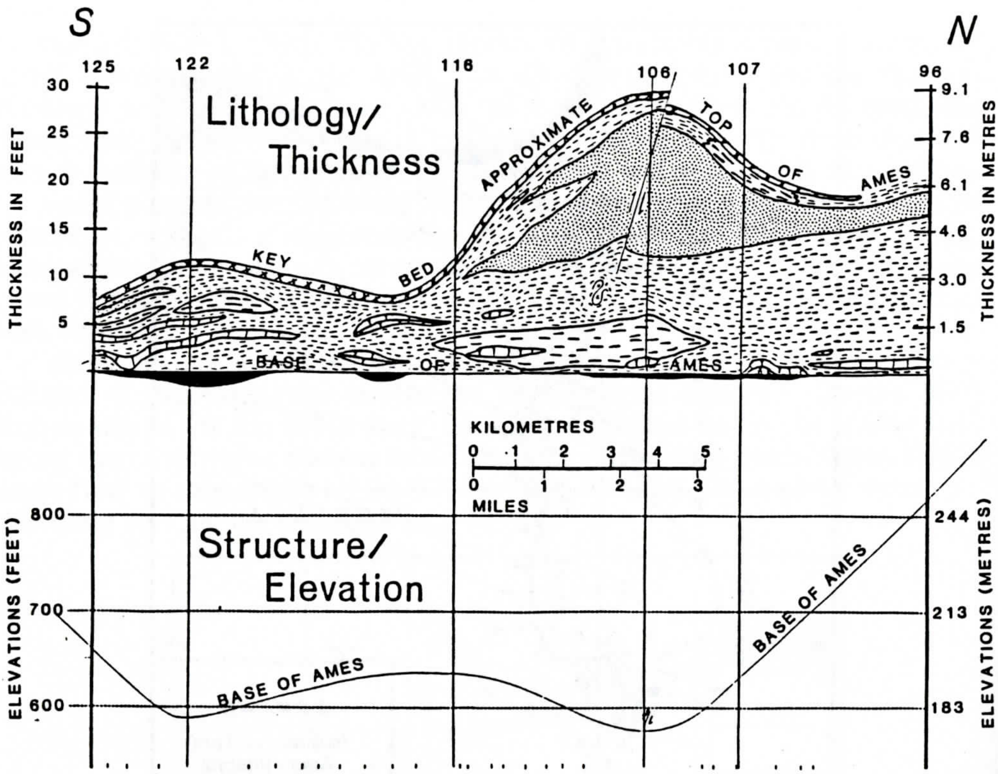

Figure 6. North-South cross-section along the Big Sandy River. Line of section runs directly between localities whose numbers are shown along the top (compare with Figure 1). Other localities used in construction of the figure are shown by tick marks along the bottom edge of the drawing. From south north these localities are: 124, 123, 98, 97, 102, 121, 119, 118, 103, 117, 104, 115, 114, 113, 112, 105, 111, 101, 146, 144, 126, 143, 142, and 141. Upper drawing uses the base of the Ames as a datum and employs conventional lithologic symbols; the heavier and more widely spaced shale symbols represent red shale (="red platy") in contrast with the lighter ones that are characteristically green- or blue-gray (="green chonetid", especially near the base), the major sandstone body is the "calcareous crinoidal sandstone" and the "key bed" shown with Xs is a sandy siderite. The "key bed", if it is non-marine, and the coal beneath the Ames are the only rocks shown that are not marine and would not be assigned to the Ames Member. Magnitude of fault displacement reduced (from 10 ft or 3 m) for clarity of presentation in the upper drawing. The lower drawing portrays the base of the Ames using elevation data. Vertical exaggeration of the upper illustration approximately 700x, of the lower approximately 70x.

south faults influenced the accumulation of fluviodeltaic deposits, including coals. The faulting progressed northward with time (Horne, 1979). The trend of the structures near Huntington parallels these fault systems, but whether they are related solely to Alleghanian "thin-skinned" tectonics or represent another reactivation of basement structures (Dever, 1986) is uncertain.

The northeast-southwest trending Pittsburgh-Parkersburg-Huntington

Syncline converges with a prominent east-west trending structure, the 38th Parallel lineament, near Huntington. Yielding and Dennison (1986) have summarized evidence for movement along the 38th Parallel lineament that may have extended from Middle Ordovician to Late Pennsylvanian time or even later. Woodward (1961) found still older structures in the subsurface along this trend, his "Lower Cambrian Scarp" and Rome Trough. Thus the study area is near the intersection of two major structural trends; the Pittsburgh-Parkersburg-Huntington Syncline, which forms the axis of the Dunkard Basin and therefore the most prominent structural feature of the Allegheny Plateau, and the 38th Parallel lineament, known to have influenced earlier Carboniferous sedimentation farther east in West Virginia. That the former clearly influenced Ames sedimentation is evident from the responses to structural trends within the Ames lithosomes as cited herein. The latter structure quite possibly influenced the more east-west deflection of the syncline's axis near its southwestern end.

SUMMARY AND CONCLUSIONS

The Ames Member differs from all older Pennsylvanian marine units in the Appalachian Basin in two important respects. First is the remarkable extent of the Ames. Most of the other named units prove to be discontinuous, heterochronous rock bodies of limited areal extent, although the parts are commonly united under a single proper name as though they were contiguous (classic example: Vanport I, Vanport II, and Vanport III of Ferm, 1970). In contrast, the Ames appears to genuinely be a single, persistent rock body of low diachroneity and a preserved extent of not less than 53,000 km^2. The older, local marine units owe their origins to local causes such as delta-switching, but a unit with the extent of the Ames implies a regional or wider controlling mechanism for its origin such as tectonism or eustasy.

The second difference lies in the origins of the non-marine rocks into which the older marine units and the Ames are intercalated. These represent prograding clastic wedges derived from the rising Appalachian Mountains. The non-marine rocks containing the lower units represent mostly deltaic or other paralic environments. In contrast, most of the non-marine environments into which the Ames sea transgressed were alluvial plain/fluvial as near Huntington (Merrill, 1986) and prior to transgression some of these must have been metres or tens of metres above sea level and at least tens of kilometres inland. Once again, a geologic event of some magnitude would be necessary to produce this relative sea level change.

Although the area currently under investigation is quite small, a mere dot on the extensive area of Conemaugh outcrop, it is located in a crucial area to draw generalizations about the nature of the Ames marine invasion basin-wide. If tectonic movement along the southern preserved extreme of the Pittsburgh-Parkersburg-Huntington Syncline admitted marine waters and determined the location of the Ames' southern shoreline, then these conclusions can be reasonably projected beyond the small study area. For nearly 100 km northeast of Huntington the Ames shoreline appears to have been located southeast of and parallel to the syncline axis, suggesting a continuation of the direct structural control for at least that far from Huntington. Northward from there, however, the shoreline turned rather sharply eastward and other controlling factors must be sought to explain the initiation and limitation of the marine inundation. Limited subsurface data suggest that faulting may have been responsible, and during a time of tectonic activity it is reasonable to seek further tectonic control for the Ames marine invasion in other parts of the basin. Whatever the specific cause, the correspondence between structural/geochronologic data from the Piedmont and the similar dates for the gentle deformation of the Ames at hand demonstrate that the intense Alleghanian deformation in the Piedmont carried across the fold belt and that the gentle tectonism had spread northwestward to near the maximum extent of deformed rocks by late in Carboniferous time.

ACKNOWLEDGMENTS

Many persons who contributed to this long-term study have already been acknowledged (Merrill, 1986). John C. Ferm deserves special acknowledgment for his long-term support of this and related studies. Charles S. Bishop obtained valuable elevation data from the Kentucky Highway Department and additional information was provided by the highway departments of all three states. Carol W. Connor discussed her findings in the Louisa, Kentucky area. Alan Saltsman read the manuscript and suggested changes. Later stages of field work (in 1986) were supported by U.S. Geological Survey Grant 14-08-0001-G1270. Ethel M. Jackson typed the many drafts of this manuscript and it was reviewed by D. T. Secor, Jr.

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ORIGIN OF THE FORT PAYNE FORMATION (LOWER MISSISSIPPIAN), TENNESSEE

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ABSTRACT

The petrology of the Fort Payne Formation was studied at four outcrops located along a line extending 260 km (160 mi) from west of Nashville to north of Knoxville, Tennessee. Chert (avg. 52 percent) is fairly uniformly distributed. Dolomite is pervasive and more common than is generally realized (avg. 30 percent). It forms two fabrics, one with rhombs disseminated in the chert (idiotopic-E) and a second of essentially continuous dolomite (idiotopic-S). Calcite averages 13 percent. Low iron calcite dominates, forming fossils, micrite and minor spar. Ferroan calcite occurs in geodes, veins, and as dedolomite. Detrital quartz is common (1 to 30 percent) in the two westernmost exposures and shale in the easternmost. Glauconite, pyrite (crystals and framboids), phosphate, sapropel, illite, void fill quartz, and ankerite are variably present in trace to a few percent quantities. The Fort Payne is dominantly dolomitic porcelanite (42 percent), with common cherty fossiliferous wackestone/packstone/grainstone (21 percent), and abundant chert and finely crystalline dolostone. The study area was part of a regional high on a broad, tectonically stable, marine ramp. This high was isolated from clastic sources to the north, northeast and east by distance and a belt of deeper water. Water depths on the high were 10 to 100 m. The Fort Payne forms the transition from the anoxic conditions represented by the Chattanooga Shale to the oxygenated shelf represented by overlying shallow marine carbonates. Dysaerobic to anaerobic bottom conditions dominated during Fort Payne deposition, but Waulsortian-type mounds developed on local, oxygenated, highs. Water depths decreased during Fort Payne time and the bottom moved up out of the oxygen-minimum zone. Opal from sponges devitrified to form the early, pervasive, chert. Dolomite formation occurred in two phases, both related to deep marine processes. The early dolomite dormed during methanogenesis; the late dolomite formed as a by-product of chert formation.

INTRODUCTION

The Fort Payne is one of the few oil and gas producing units in Tennessee hence knowledge of the pattern and character of its lithologic variation is of economic interest. The purpose of this paper is to present petrologic data for the Fort Payne in northcentral Tennessee as a necessary background for interpretation of both the environmental conditions under which it was deposited and its diagenetic history. To facilitate understanding the Fort Payne will be compared to the Miocene Monterey Formation.

The Fort Payne Formation is a lower Mississippian dolomitic porcelanite that caps the western Highland Rim, caps or is in the shallow subsurface in the eastern

Highland Rim, and is present in the subsurface throughout the Cumberland Plateau (Figure 1). It is similar to a number of coeval units over a very large area (Gutschick and Sandberg, 1983). In Tennessee and to the south the Fort Payne maintains its cherty character, and facies variations are on a broad scale (Milici and others, 1979; Thomas and Cramer, 1979; Thomas, 1979). More complex conditions exist to the north where it interacts with clastics of the Borden delta (Rice and others, 1979; Lewis and Potter, 1978). It is coeval with a portion of the Arkansas Novaculite (Gutschick and Sandberg, 1983). In most exposures megascopic nodules and beds of black to brown chert in the gray to brown porcelanite background make the Fort Payne easy to identify. In some areas and intervals the Fort Payne has a shaley appearance. Local crinoid/bryozoan buildups similar to Waulsortian-type mounds are common (MacQuown and Perkins, 1982). Its conformable contact with the underlying Chattanooga Shale varies from sharp to transitional. The contact with the overlying Warsaw Formation is difficult to pick but is apparently conformable. As established below, it forms the transition from the starved anoxic shelf represented by the

	FORMATION	THICK.	LITHOLOGY
MISSISSIPPIAN SYSTEM	PENNINGTON FM.	150 To 450 Ft.	SHALE AND SILTSTONE (RED AND GREEN), SANDSTONE, DOLOSTONE AND LIMESTONE.
	BANGOR LS.	150 To 300 Ft.	LIMESTONE, FOSSILIFEROUS, OOLITIC IN PART.
	BIG CLIFTY FM.	0 To 60 Ft.	SANDSTONE, SANDY LIMESTONE, FOSSILIFEROUS SHALE, DOMINANT LITHOLOGY IS VARIABLE.
	MONTEAGLE LS.	150 To 275 Ft.	LIMESTONE, FOSSILIFEROUS, OOLITIC & CROSS-BEDDED IN PART.
	ST. LOUIS LS.	MAX. 110 Ft.	LIMESTONE & DOLOSTONE, CHARACTERIZED BY LITHOSTROTION CORAL.
	WARSAW LS.	MAX. 80 Ft.	LIMESTONE, ARENACEOUS, CROSS-BEDDED IN PART.
	FT. PAYNE FM.	MAX. 190Ft.	LIMESTONE AND DOLOSTONE, VERY SILICEOUS.
DEV.?	CHATTANOOGA SH.	0 To 40 Ft	SHALE, DARK GRAY TO BLACK, BITUMINOUS, FISSILE.

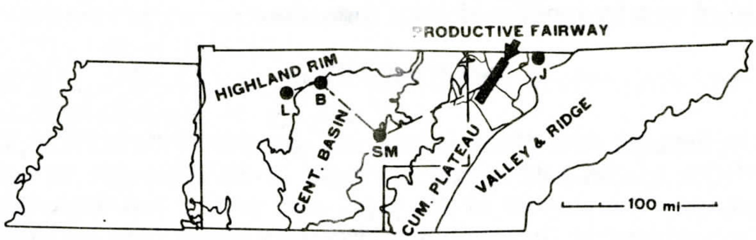

Figure 1. A - Mississippian stratigraphic terminology used in Tennessee (Stearns, 1963). The term Big Clifty was reintroduced by Roberts and Lumsden (1982). B - Outcrop location map. L - Lillimay, B - Bethel, SM - Short Mountain, and J - Jellico. The productive fairway marks the general trend of oil production in the Fort Payne Formation.

Chattanooga Shale to the aerobic shallow marine carbonate shelf represented by the Warsaw or St. Louis Formations. Its diagenetic history involves deep subtidal formation of dolomite associated with chert and organic alterations.

METHODS

The petrology of the Fort Payne Formation was studied at three outcrops in central Tennessee (Anderson, 1981) and at one in the eastern part of the state (Gregory, 1981). These complete sections form a line extending 260 km (160 mi) from west of Nashville to north of Knoxville, Tennessee (Figure 1). Locations were chosen for their complete exposure and the presence of top and bottom contacts. Numerous partial exposures were also examined. Field and hand specimen study supplemented data obtained from analysis of 142 alizarin red and potassium ferricyanide stained thin sections. Slides were either point counted or compared to visual standards to determine mineralogy (detrital quartz, authigenic chert, dolomite, calcite) and fabric (carbonate grains, micrite, and spar). Insoluble residue analysis using HC1 was performed on most samples. X-ray diffraction was used to verify gross composition, the presence and composition of clay minerals, and the degree of nonstoichiometry of the dolomite. SEM and luminescence petrography were used to examine textural details.

The percent of noncarbonate material determined by point count did not correlate well with the total percent insoluble residue. In order to obtain consistent comparable data, the proportions of the four major components (chert, dolomite, calcite, detrital quartz) were determined by apportioning the amount of dolomite and calcite determined in thin section into the percent soluble component of the rock. This procedure mixes weight and volume percent data and includes all non- detrital quartz and other insoluble material with the chert. The discussion that follows emphasizes general trends and does not depend upon variations of less than several percent.

Six lithologic groups are recognized. One group can be described using the wackestone to grainstone terms of Dunham (1962). Five additional groups are based on the relative amount of chert, dolomite and calcite, modified by textural considerations. The six groups are described under results.

RESULTS

Mineralogy (Figure 2)

Chert forms a grand average of 52 percent of the Fort Payne. It is mostly in dolomitic porcelanites but also in relatively pure megascopic nodules, geodes, and separate beds. In thin section the chert is overwhelmingly fine (about 0.005 mm) and equigranular. Fibrous chalcedonic chert forms abundant millimeter to decimeter scale nodules. Length-fast chalcedony dominates length slow. Samples treated with HC1 typically retain their original size and shape, and bedding is continuous from unweathered outcrops into adjacent saprolith. This suggests that chert forms a continuous phase on both small and large scales.

Dolomite averages 30 percent and is the dominant carbonate mineral at all four locations. Dolomite rhombs have a bimodal crystal size distribution. The smaller mode crystals are clear, unzoned, equidimensional, vary from 0.01 to 0.05 mm in size (mode 0.020 to 0.025 mm), and occur as isolated individuals embedded in a chert background. They luminesce a uniform pink. Some of these crystals are fractured. The contact between the rhombs and surrounding chert is sharp. The rhombs resemble the isolated crystals common in deep marine sediments (Lumsden, 1983), albeit somewhat larger on the average. This idiotopic-E texture grades into idiotopic-S (Gregg and Sibley, 1984) where larger crystals coalesece into an interlocking mosaic of dolomite with subordinate chert. The dolomite has a luminescent core (in the same size range as that of the smaller dolomite crystals) enclosed by a broad nonluminescent band that is occasionally, in turn, rimmed by a thin, dull-luminescent band. This relationship suggests later growth on a luminescent core. Dolomite at Jellico is cloudy centered xenotopic-A with a nonluminescent core an luminescent rim. Ankeritic dolomite (nonluminescent dolomite that stains a pale blue) forms large isolated rhombs and fills some fractures and voids. The [104] X-ray peak for dolomite is shifted, i.e. it is nonstoichiometric (mode 55-56% $CaCO_3$). The presence of two phases of dolomite and ankerite prevents interpretation of this characteristic.

Calcite averages 13 percent. Low iron calcite forms fossils, micrite and uncommon void fill spar. Ferroan calcite fills fractures, vugs and quartz geodes and partially to completely replaces dolomite rhombs (dedolomite).

Quartz forms 1 to 30 percent fine sand size, angular, single crystal detrital grains at the two westernmost exposures (Figure 2). Replacive quartz with

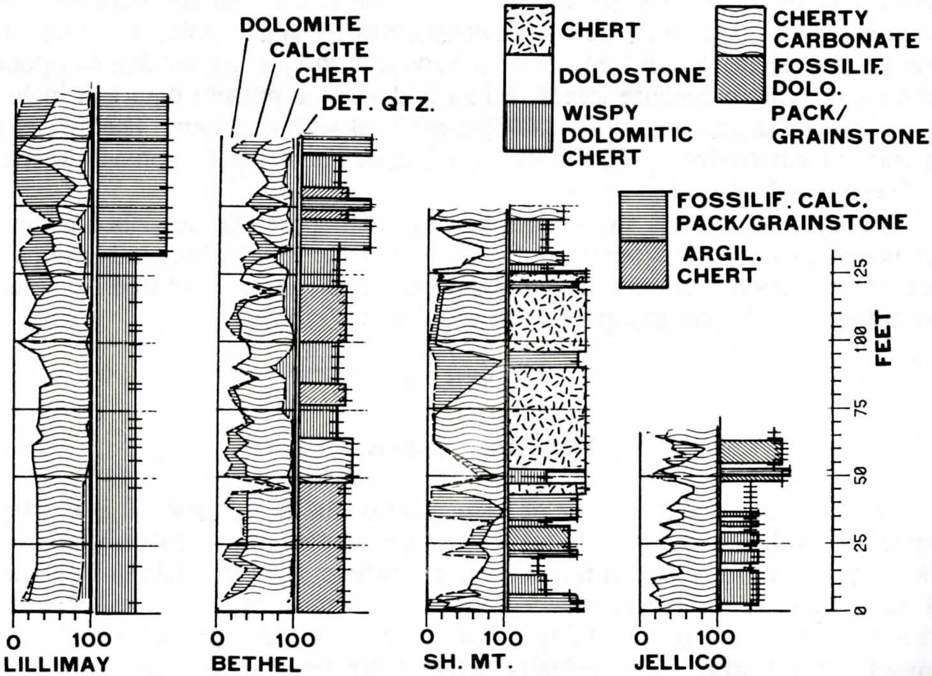

Figure 2. Mineralogical variation (left) and generalized columnar sections (right) of the Fort Payne. The "weathering profile" is arbitrary and intended to highlight lithologic variations. The common shale units at Jellico are to thin to show. Tick marks on the right margin mark sample points.

flamboyant brush extinctions and complex intercrystalline boundaries forms nodules and geodes.

Glauconite forms sparse minute grains, typically less than 0.05 mm in diameter and less than 1 percent of the area of a thin section. It replaces some sponge spicules at Jéllico. It is common where clays are present, and absent where packstones are common.

Phosphatic and sapropel material as blebs, pellets and uncommon fossil fragments occur sporadically, but seldom exceed one percent of a slide. Marcher (1962, p. 822), Lineback (1966, p. 23) and Bassler (1932) variously report organic carbon, carbonaceous and sapropelic material and megascopic phosphate nodules.

Anhydrite forms rare minute crystals in some chert nodules (see also Chowns and Elkins, 1974).

Illite is disseminated in chert at the base of the Bethel and Short Mountain sections. Illitic shale beds and stringers form 18 percent of the Jellico section.

Pyrite forms trace to two percent quantities in the Fort Payne. It forms minute framboids at Lillimay and Bethel, and forms crystals visible in thin section at Short Mountain. It is absent at Jellico.

Lithofacies (Figure 2)

Dolomitic porcelanite is the most abundant lithology (42 percent of aggregate measured section). In thin section it has a characteristic appearance with wispy stringers of dolomite rhombs and dolomicrite running through a background of finely crystalline chert (Figure 3). Sponge spicules are common. It appears to be

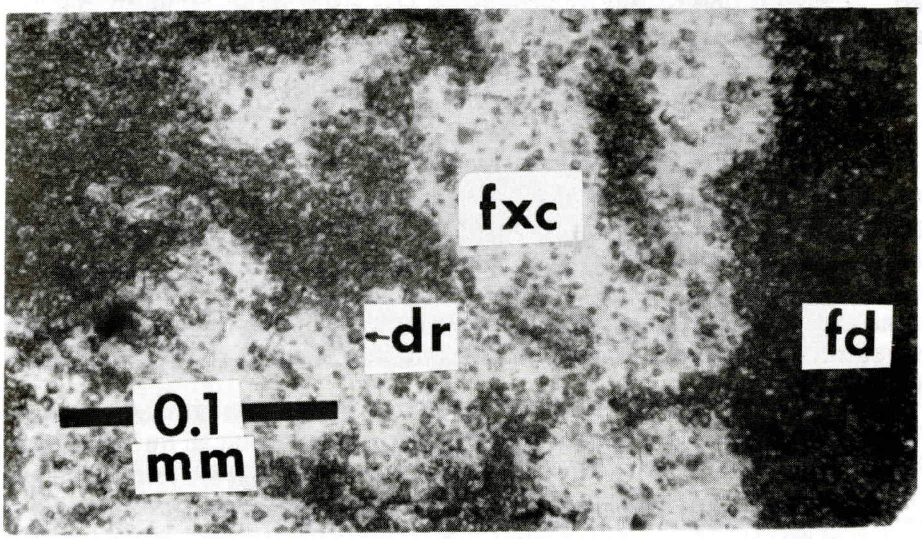

Figure 3. Thin section photomicrograph (uncrossed nicols) of the common wispy texture of the dolomitic porcelanite. The light areas are finely crystalline chert (fxc); the dark areas are a mosaic of fine dolomite crystals (fd). Scattered dolomite rhombs (dr) are present in the chert. The sample is from 3 m above the base of the section at Jellico.

a chert and dolomite replaced lime mudstone or wackestone. It is dominant at Lillimay and common elsewhere.

Cherty fossiliferous wackestone/packstone/grainstone (21 percent) has abundant primary calcite, largely as fossils but also as micrite and minor spar cement. It corresponds to the micrite charged crinoid/bryozoan Waulsortian-type mounds described by MacQuown and Perkins (1982). This group has very little chert.

Chert samples (12 percent) have 80 percent or more insoluble residue. Ghosts of the original fabric are common, and most cherts appear to be replaced micrite, fossiliferous wackestones or packstones. Isolated dolomite rhombs and partially altered calcite fossils are enclosed in the chert. This group is common only at Short Mountain.

Dolostones (5 percent) have an interlocking mosaic of dolomite crystals (idiotopic-S). Chert is present but is masked by the dolomite in thin sections. Dolostones are most common at Jellico.

Cherty carbonates (5 percent) include calcareous cherts and carbonate mudstones with variable amounts of calcite along with chert and dolomite. They represent different degrees of chertification of a mixed calcite/dolomite mudstone.

Argillaceous chert (15 percent) is a blur of clays, fine carbonate and chert with traces of muscovite. It forms the basal 50 ft (16 m) of the Bethel section.

Fauna

The dominant fossils are sponges, represented by needle-like forms in the chert. Crinoids, bryozoans, brachiopods and ostracodes occur in mounds of wackestone and packstone.

Sedimentary Structures

Bedding is parallel, continuous and thin (1-5 cm). In unweathered exposures beds appear thicker (25-30 cm). Low relief (10 cm to 1 m) Waulsortian-type mounds are sporadically present (Figure 4). Rare small scale cross beds and slump features occur. On a smaller scale (1 cm) a profusion of horizontal mottles are present, assumed to be burrows (Figure 4). No vertical burrows or sedimentary structures produced by waves, currents or exposure to the atmosphere were seen. Chert nodules, quartz and calcite lined geodes and vugs are abundant (Figure 4).

Thicknesses

The Chattanooga Shale thins onto the Nashville Dome and Cincinnati Arch, suggesting that these were relatively positive, but nevertheless submarine, features during its deposition (Craig and Varnes, 1979). Although thickness variation for the Fort Payne is not as well documented, available data suggest the same pattern and conclusion. Fort Payne thickness is fairly constant in central Tennessee (58 m at Lillimay, 53 m at Bethel, 45 m at Short Mountain) but thins to 21 m at Jellico (Figure 2). Milici and others (1979) report thicknesses of 30 to 60 m over the study area. MacQuown and Perkins (1982) show logs with approximately 40 m

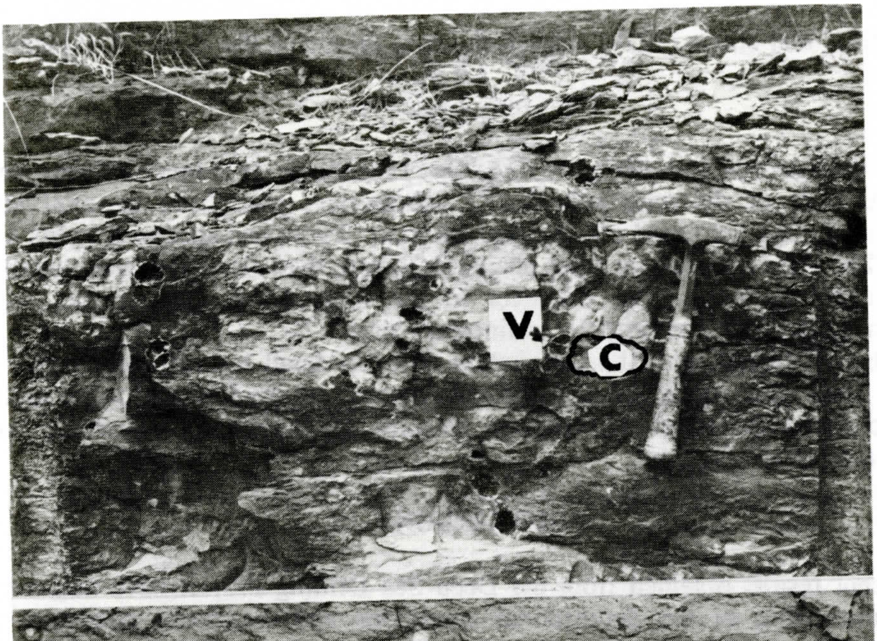

Figure 4. A - A crinoid/bryozoan packstone mound at Lillimay with approximately 15 cm of relief on its upper surface. Several quartz lined vugs, up to 5 cm long are present (V). Chert nodules are common (C). The hammer is 32.0 cm (12.5 in) long. B - A closer view of non-mound sediments at Lillimay. Note the abundant evidence of horizontal bioturbation. Note also the abundant small (approximately 1.0 cm), white quartz/calcite nodules. The lense cap is 5.0 cm (2.0 in) in diameter.

of Fort Payne in the productive fairway (indicated in figure 1). To the northwest. and northeast the Fort Payne thickens into the basin that separated the study area from the Borden delta (Lineback, 1966; Lewis and Potter, 1978; Cluff, 1980) and also thickens north into the Cumberland Saddle. To the east, into the Appalachian

Basin, Milici and others (1979) show it thinning on top of the Borden/Grainger.

The Study Locations

Jellico contrasts with the other three study locations. It is much thinner, and has more beds of dolomite and shale. Glauconite, phosphatic material and disseminated clays are found only toward the base. Pyrite and sponge spicules are absent. The chert nodules are more chalcedonic here than elsewhere and contains no anhydrite vestiges. The character of the dolomite is also distinct with nonluminescent cores and brightly luminescent rims.

Lillimay, Bethel, and Short Mountain differ in detail but are comparatively similar. They have similar thicknesses and common wackestone/packstone units. Pyrite, phosphate/sapropel and sponge spicules are common; glauconite and ankerite are variably present.

DISCUSSION

Recent results obtained from the Deep Sea Drilling Project have improved our understanding of the depositional conditions and diagenetic pathways of upwelling influenced deep water carbonate - siliceous - organic rich deposits (Cook and Enos, 1977; Pisciotto, 1981; Arthur and others, 1984; Baker and Burns, 1985). This knowledge has been used in studies of the Miocene age Monterey Formation of the southern California coastal area; both from a general point of view (Garrison and others, 1981; Graham and Williams, 1985) and with regard to the origin of its dolomite (Garrison and others, 1984; Burns and Baker, 1987). Both the Fort Payne and Monterey have abundant dolomitic porcelanites associated with organic material and the dolomite character in the Fort Payne is similar to that of the Monterey (e.g. Isaacs, 1984). The stratigraphic similarity of the two deposits is enhanced if one considers the Chattanooga Shale and Fort Payne as a depositional unit, thus providing the Fort Payne with the clastics and abundant organic material observed in the Monterey. As the Monterey is a significant oil and gas producer (Crain and Mero, 1984) it forms an interesting model for the Fort Payne.

Tectonic Setting

The Fort Payne/Chattanooga and Monterey differ sharply in tectonic setting. The Chattanooga was deposited over a vast area of stable craton during a rapid transgression onto a low relief, generally karst, terrane. The Fort Payne and coeval units form an isochronous interval (anchoralis-latus Zone of Gutschick and Sandberg, 1983) deposited conformably on the Chattanooga. Clastics are minor in the Fort Payne. In contrast, the Monterey was deposited over a much smaller area on a tectonically unstable shelf divided into numerous blocks and basins with considerable clastic input from a nearby continental source (e.g. Graham and Williams, 1985).

Within the study area, the Fort Payne was deposited on a regional high on a deep passive margin, part of the Fort Payne Ramp of Gutschick and Sandberg (1983). Thickness variations suggest that the study area was isolated from the clastics of the Borden Delta to the northwest, north, northeast, east, and southeast

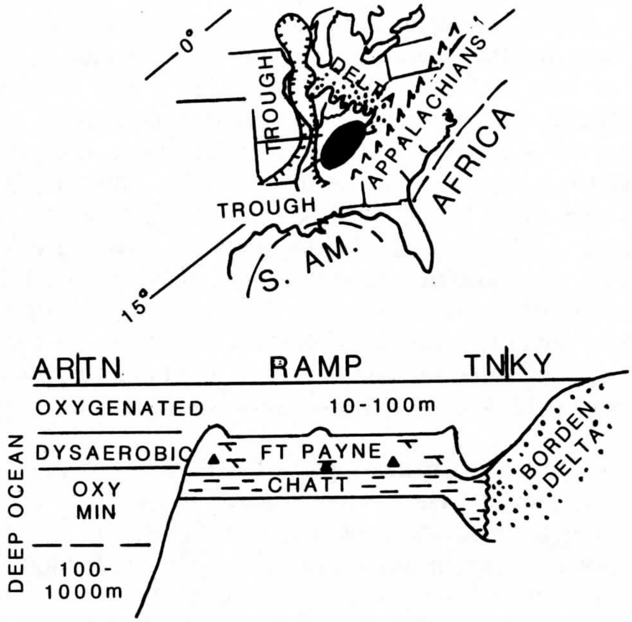

Figure 5. Schematic representation of relationships during Fort Payne deposition. A - map view. B - diagramatic cross section of A. After various sources including Lineback (1966), Lowe (1975), Lewis and Potter (1978), Sedimentation Seminar (1981), Gutschick and Sandberg (1983), and Arthur and others, (1984).

by a belt of relatively deeper water (Figure 5). To the west both Lowe (1975) and Gutschick and Sandberg (1983) suggest that a silled deep water trough, connected to the open sea, extended north parallel to the trend of the modern Mississippi River. To the south conditions are uncertain (Thomas, 1979; Thomas and Cramer, 1979); however, Gutschick and Sandberg (1983) suggest that a deep marine trough separated the study area from the Mississippian age equivalent to South America.

Deposition

The lack of precise dates for the start and end of Fort Payne deposition limits the accuracy of estimates of the average rate of sedimentation. Using the time scale of Palmer (1983) and assuming deposition coincident with the Vla stage of the Visean, the Fort Payne represents approximately 2 My of time. Gutschick and Sandberg (1983) suggest 1.5 My for duration of the anchoralis-latus Zone, an interval including deposition of the Fort Payne. Taking 40 to 60 meters as representative thicknesses gives average rates of sedimentation in the range of 20 to 40 m/My (undecompacted). This is similar to the 40 to 90 m/My (also undecompacted) for the carbonate and siliceous facies of the Monterey (Burns and Baker, 1987), and much greater than the 4 m/My for deep marine sedimentation (calculated from an approximate average of 1 g/cm^2/ 1000 yrs in figure 3 of Davies and Worsley, 1981).

The story of Fort Payne deposition begins with that of the Chattanooga. This laminated, black, petroliferous, pyritic, quartz-dominated shale was deposited at a low sedimentation rate on a bottom in the anoxic-anaerobic interval. Conant and Swanson (1961) state that the Chattanooga Shale was deposited in "relatively quiet water at depths of tens of feet". As the study area was at approximately 15 degrees south latitude at the time of Chattanooga deposition (Gutschick and Sandberg, 1983), such a shallow water area should have been the site of high carbonate productivity. As this was not the case, the water depth was probably below the photic zone. The fine grained nature of the Fort Payne, the absence of sedimentary structures produced by shallow water or subaerial environments, the typical uniform and thin bedding, the dominance of horizontal trace fossils, the nature and limited diversity of the fauna, the low sedimentation rate, and the presence of organic material combine to suggest deposition on a subtidal shelf in water below the depth of wave and strong current agitation. It follows that ten to a hundred meters would be a good generalization for the water depth of the Fort Payne.

The transition from the Chattanooga to the Fort Payne came about through the interaction of two factors. The relative position of the sea floor moved from its location in the oxygen minimum zone into the dysaerobic interval. This was accompanied by an increase in silica and carbonate production; the former as sponges, the latter principally in Waulsortian-type mounds on local highs and along the platform margins. Sedimentation was dominated by the proliferation of sponges fed by silica supplied by upwelling waters from open ocean troughs to the west and southwest (Lowe, 1975). The Waulsortian-type mounds, while not shallow water features (Wilson, 1975, p. 166), extended up into oxygenated waters, and the micrite produced on these banks was disseminated into adjacent areas. This micrite provided the basis for later dolomite formation. Overall the water became more shallow as Fort Payne deposition progressed. Whether this was due to a fall in sea level, tectonic uplift of the ramp, buildup of the sea floor by sedimentation, or a combination of factors is not clear.

In contrast to the deep water hypothesis supported by data in this study Chowns and Elkins (1974) proposed a shallow water evaporite environment. This was necessary for explaining the abundant quartz nodules and geodes as replacements of anhydrite in the Fort Payne. Schmalz (1969) proposed a deep water model for evaporite formation however this does not offer a compromise for the Fort Payne as any deep brines would flow into the deeper water surrounding the study area. Nor can we suppose the nodules are due to local conditions; they are ubiquitous.

Dolomite Formation

The alteration of the original mix of biogenic opal and calcite micrite to the dolomitic porcelanite of today is a complex story only briefly summarized here.

Chert formed in two, possibly three, phases. The first and quantitatively overwhelming phase, the fine equigranular chert, formed by mobilization of silica in sponges that then replaced the original opal and primary calcium carbonate (Marcher, 1962; Chowns and Elkins, 1974). Evidence includes chert replaced sponge spicules, partial to complete replacement of carbonate fossils, ragged-transitional boundaries between patches of chert and relatively unreplaced fossiliferous wackestones/packstones, and similarity in textures between chert and carbonate areas. A second, much later, chert phase consists of fibrous chalcedonic chert and quartz that form geodes and nodules (Chowns and Elkins, 1974). These chert types also line some fractures suggesting a third, post fracture, mobilization of quartz.

Dolomite also formed in two phases. The similarity in size and appearance of the smaller (entirely luminescent) rhombs to those observed in most modern lake, supratidal and deep marine sediments implies that this dolomite formed early, probably as a diagnetic product in the newly deposited sediment. A second, pervasive, phase of dolomitization is represented by the common mosaics of larger dolomite crystals. Some of these large crystals encroach on the margins of detrital quartz grains and therefore must have formed after the detrital quartz was deposited.

Models of dolomite formation based on studies of modern supratidal carbonates are clearly inadequate for the Fort Payne in view of the overwhelming evidence for its deep subtidal origin in the study area. There is no relation between the amount of dolomite (and chert) and formation contacts; therefore the mechanism proposed by Knauth (1979), where silicification and dolomitization are caused by an influx of external fluids, also does not appear suitable. Clay diagenesis as a dolomite (McHargue and Price, 1982) and chert (Strom and others, 1981) source is unlikely as there is no relationship between chert/dolomite abundance and the presence of underlying and interbedded shale (Figure 2).

Dolomite formation at Lillimay, Bethel, and Short Mountain can be explained by a combination of two models proposed by Baker and his coworkers. Early dolomite formation was controlled by microbial sulfate reduction in the upper few meters of sediment (Baker and Burns, 1985). This model can account, in part, for the relatively small amount of dolomite encountered in DSDP smaples, and the similar fine sized (early?) dolomite in the Fort Payne. Dolomite formation is inhibited by the transformation of opal-A (amorphous silica) to opal-CT, but the subsequent transformation of opal-CT to chert favors dolomite formation (Baker and Kastner, 1981). Thus the second, pervasive, phase of dolomite formation and the formation of nonluminescent rims on luminescent cores followed the opal-CT to chert transformation and depended on Mg release by that alteration. This late dolomite formed at the expense of calcite, as suggested by the inverse relationship between the abundance of calcite and dolomite in going from west to east (cf. Isaacs, 1984). Dolomite at Jellico formed differently, just how is uncertain.

CONCLUSIONS

The Fort Payne is dominantly dolomitic porcelanite (42 percent) with common cherty fossiliferous wackestone/packstone/grainstone (21 percent), and abundant chert and finely crystalline dolostone. The study area was part of a regional high, isolated from clastic sources to the north, northeast and east by a distance and a belt of deeper water. This high was part of a broad, tectonically stable, marine ramp where water depths ranged from 10 to 100 m. The Fort Payne forms the transition from the anoxic conditions represented by the Chattanogga Shale to the oxygenated shelf represented by overlying shallow marine carbonates. Dysaerobic to anaerobic bottom conditions dominated during Fort Payne deposition, but Waulsortian-type mounds developed on local, oxygenated, highs.

Water depths decreased during Fort Payne time and the bottom moved up out of the oxygen-minimum zone. Opal from sponges devitrified to form the early, pervasive, chert. Dolomite formation occurred in two phases, both related to deep marine processes. The early dolomite formed by sulphate reduction; the late dolomite formed as a by-product of chert formation.

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MULTIPLE EVENT STRATIFICATION IN CARBONATE INTRACLAST CONGLOMERATES IN THE CAMBRIAN OF SOUTHWESTERN VIRGINIA

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ABSTRACT

Petrographic and paleocurrent data from three multiple event stratification sequences within the Copper Ridge Formation (Upper Cambrian) in southwestern Virginia are presented to illustrate the importance of such information in interpreting depositional environments. The event sequences are found in storm-generated carbonate conglomerates containing abundant imbricate intraclasts. The first succession consists of two events: a lower zone of parallel-bedded cryptalgal laminites with intercalated intraclasts of the same material and an upper zone of more heterogenous intraclasts and interstitial grains. Both events occurred in an intertidal setting where storm-enhanced tidal currents were operative. The second sequence is another double event occurrence in which both intervals contain varied types of intraclasts and interstitial grains (ooids, peloids, bioclasts, small intraclasts) but more mud is present in the upper unit. These sequences originated as resedimented subtidal shoal deposits over which clasts from muddier environments were driven subsequently. The third event succession consists of three or four mixed siliciclastic-carbonate subunits. Each subunit contains mudstone intraclasts, some showing shrinkage cracks, overlain by rippled quartz arenite caps. The subunits formed in an intertidal setting in which desiccated mud chips were reworked by storms approaching from various quarters. Waning currents deposited quartz-rich sands at the close of each high-energy event.

This study indicates that multiple event stratification is common in peritidal carbonate conglomeratic intervals and that significant details of environmental interpretation can be lost if such stratification is not recognized and properly analyzed.

INTRODUCTION

Event stratification may be thought of as a sedimentary unit possessing distinctive depositional and erosional features caused by events of significantly greater magnitude than those operative during ordinary day-to-day sedimentation. Although such events may be "rare" in terms of human life span and experience, deposits formed by them are common in the stratigraphic record (Seilacher, 1982). Major storm sequences ("tempestites") are illustrations of sediments resulting from such events.

During the course of a comprehensive analysis of intraclast conglomerates in some ancient platform carbonates in southwestern Virginia (Whisonant, in press), a number of excellent examples of multiple event stratification were observed in these coarse-grained units. The purpose of this paper is to present petrographic and paleocurrent data from three such occurrences. Such information is significant because little work is available concerning multiple event sequences, particularly in shelf carbonates (see, e.g., Einsele and Seilacher, 1982) and studies concerning shallow marine conglomerate fabrics are also very limited (Bourgeois and Leithold, 1984).

The intraclast conglomerates described are from the Copper Ridge Formation which is Late Cambrian in age and constitutes part of the three km-thick Lower Paleozoic passive margin carbonate sequence within the Appalachian miogeocline. The Copper Ridge crops out extensively in belts in the northwest part of the Valley and Ridge Province in southwestern Virginia (Figure 1). Here it is about 400 m thick and contains much dolomite with lesser amounts of quartz sandstone (Koerschner, 1983). A Copper Ridge stratigraphic equivalent, the Conococheague Formation, was deposited more seaward (southeastward), is thicker, and contains abundant limestone as well as dolomite and minor quartz sandstone compared to the Copper Ridge (Markello and Read, 1982). Previous studies of both formations in the study area interpret them as marine shallow water peritidal carbonate deposits (Markello and others, 1979; Read, 1983; Koerschner, 1983; Whisonant and others, 1984; Whisonant and Maloney, 1985). A depositional strike of N 40° E (present coordinates) for the Late Cambrian shelf in southwestern Virginia has been determined by Markello and Read (1982, p. 874) from lithofacies patterns and is supported by the more regional paleogeographic maps of Laporte (1971, p. 729) and Scotese and others (1979, p. 237).

The Copper Ridge contains numerous intraclast conglomerates. Such layers are particularly abundant in Lower Paleozoic carbonates in various parts of the world (Markello and Read, 1981; Sepkoski, 1982). Although a number of distinctive types and origins of intraclast strata in the Copper Ridge and associated units have been described (Markello and Read, 1981; Demicco and Mitchell, 1982; Koerschner, 1983; Whisonant, 1983; Kozar and others, 1986), this study

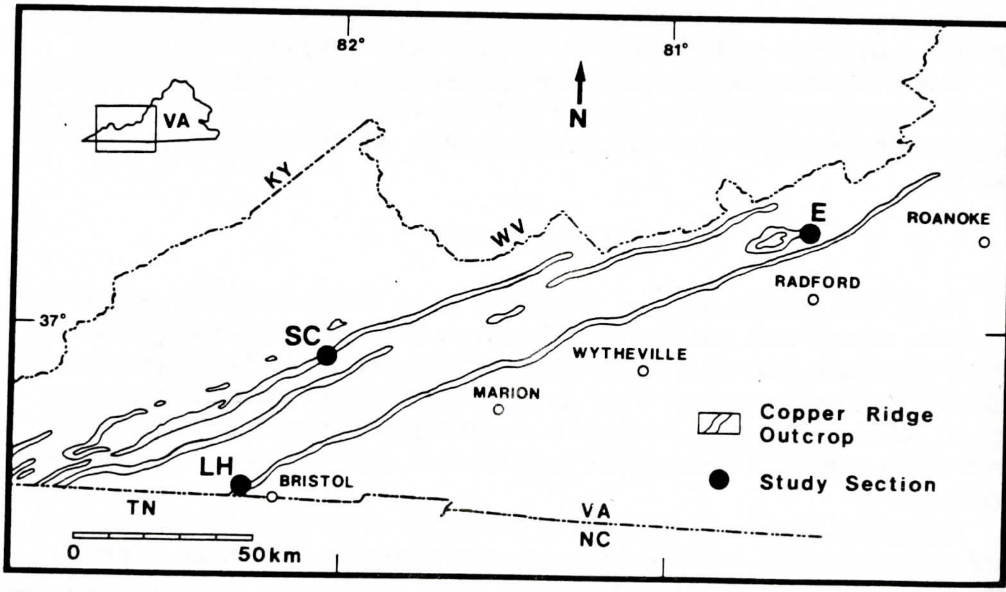

Figure 1. Index map showing Copper Ridge outcrops and study sections in southwestern Virginia. Study sections are E - Eggleston, LH - Lime Hill, SC - Spring City.

deals only with event sequences in those conglomerates containing abundant imbricate intraclasts because of their utility in paleocurrent analysis. These and similar intraclast beds in the study formation are storm-generated (Markello and others, 1979; Demicco and Mitchell, 1982; Koerschner, 1983; Whisonant, 1983; Whisonant and others, 1984).

METHODS

Distinctive examples of multiple event sequences in intraclast conglomerates were selected for detailed study at Eggleston, Lime Hill, and Spring City, Va. (Figure 1). Oriented blocks containing abundant imbricate intraclasts were analyzed in the laboratory for paleocurrent and petrographic information. Directional data were obtained by sawing faces in the sample blocks so that two planar surfaces normal to bedding and approximately normal to each other were available to observe fully the planes of clast imbrication. A board was oriented in these planes from which the azimuths of the maximum dip directions were measured. Only discoidal or blade-like clasts, which constitute the vast majority of the intraclasts, were used because (1) the least equidimensional clasts in conglomerates are the most likely to show any preferred orientation developed during transportation and deposition (Hendry, 1976) and (2) flow direction variability has been found to increase if conglomerate grains of all shapes are used in fabric analysis (Potter and Pettijohn, 1977). Those clasts larger than others in the population, generally > 2 cm, were selected to avoid the problem of smaller clasts assuming orientations due to settling in the spaces among the larger clasts (Davies and Walker, 1974; Hendry, 1976).

All imbrication data are plotted into 30° sector rose diagrams. These are presented so as to indicate direction of current flow, i.e., readings are plotted 180° from the actual azimuths of the imbrication plane dips.

Polished and acid-etched slabs as well as thin sections were prepared from all event sequences for petrographic analysis.

MULTIPLE EVENT STRATIFICATION: PETROGRAPHY AND PALEOCURRENTS

Eggleston

The first multiple event zone discussed is part of a Copper Ridge section located at Eggleston, Va. (Figure 1). Koerschner (1983) has described this exposure in detail.

The event sequence consists of two completely dolomitized conglomeratic units (Figure 2). The lower event is represented by a 22-25 cm-thick sequence of parallel-bedded dolomud cryptalgal laminites reworked into ungraded intraclast intervals in places, particularly in the upper part. The parallel-bedded material shows abundant desiccation cracking and laminoid fenestrae. The intraclasts, which are internally laminated dolomud or peloidal dolomud, show these features also. In addition, many intraclasts are bent, reflecting an original flexibility typical of algal sediment. The intraclast shapes are commonly blocky and angular, suggesting a minimum of transport during reworking. The conglomerates are typically clast-supported with interstitial grains mainly dolosand-silt and lesser amounts of small intraclasts and peloids. Mud-floored shelter voids are present beneath some large intraclasts.

The lower zone is capped (where not erosionally truncated by the upper unit) by a ripple-bedded dolosand-silt grading into dolomud at the top. Intraclast imbrication in this lower zone shows considerable scatter, but clearly indicates general shoreward movement toward the northwest (Figure 2). The bimodality seen in the rose diagram might reflect flood-ebb flow.

The upper event is represented by an 8-10 cm-thick unit that channels into the lower sequence and contains more heterogenous intraclasts and interstitial grains (Figure 2). This conglomeratic interval is ungraded to crudely inversely graded; indeed, the largest clast observed in the sequence is very near the top. The intraclasts are supported by a matrix composed of dolosand-silt, ooids, peloids and small intraclasts. The intraclasts are more rounded than those in the underlying event and are a mix of structureless dolomud, laminated dolomud, and wackestone (some oolitic) with lesser amounts of packstone.

Paleocurrents deduced from intraclast imbrication in this upper zone show a well-developed bimodal-bipolar pattern that is oriented approximately normal to the shoreline (Figure 2). This strongly suggests a tidal current influence, perhaps enhanced by storm activity. Note also in Figure 2 that combining directional data from both lower and upper events results in significant information loss. The combined rose diagram is polymodal and suggests general transport toward the northwest; however, the individual event patterns, crucial to correct environmental reconstruction, are obscured. This same problem of losing important details of paleocurrent activity by combining directional readings is present in the other multiple event sequences discussed below.

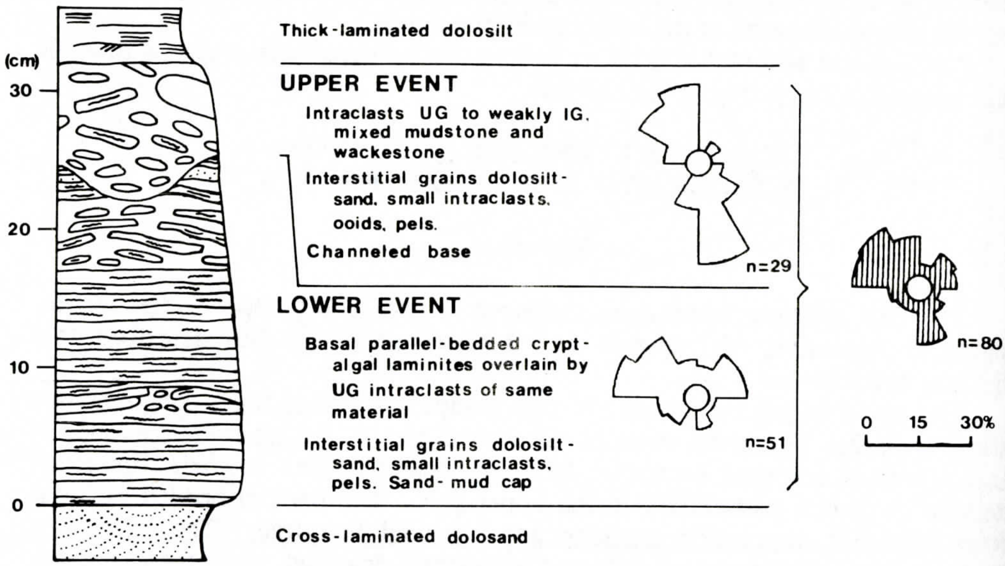

Figure 2. Event stratification at Eggleston. Grading symbolism for this and succeeding figures is UG - ungraded, IG - inversely graded, NG - normally graded.

The lower event is interpreted as representing an intertidal algal mudflat environment over which storm waves broke and reoriented some of the laminated mud intraclasts produced by prior exposure and desiccation. The intraclasts evidently were carried in a net landward direction, although very little actual movement seems to have occurred. The graded cap suggests that waning traction transport after the uppermost storm event first deposited ripple-bedded sand-silt; mud then settled from suspension to complete the sequence.

Following deposition of the lower unit, a very shallow tidal channel formed in this intertidal setting in which intraclasts and interstitial grains from more varied environments were transported. This event does not reflect ordinary tidal current activity, however, because the intraclasts are ungraded to inversely graded and matrix-supported. Rather, this is probably a high-energy storm occurrence enhancing tidal currents to create a flow regime in which both turbulence and dispersive pressures were important in transporting the large clasts (Pierson, 1981). Similar flow dynamics inferred for one of the Lime Hill event units are discussed in more detail below.

Overlying this upper event is a parallel-to ripple-bedded thick laminated dolosilt that represents a return to more ordinary day-to-day sedimentation in a peritidal environment.

Lime Hill

The next multiple event succession is within the Lime Hill section (Figure 1). This exposure is particularly interesting because it occurs in the southeasternmost outcrop belt and is therefore much limier than most Copper Ridge sequences. Thus, dolomitization has not obliterated many of the primary depositional features as occurs in most of the Copper Ridge.

The event subunits consist of two amalgamated conglomerates containing rounded to well-rounded intraclasts (Figure 3). Beneath the lower subunit is a cross-laminated, well-sorted grainstone consisting nearly entirely of medium sand-sized ooids and intraclasts.

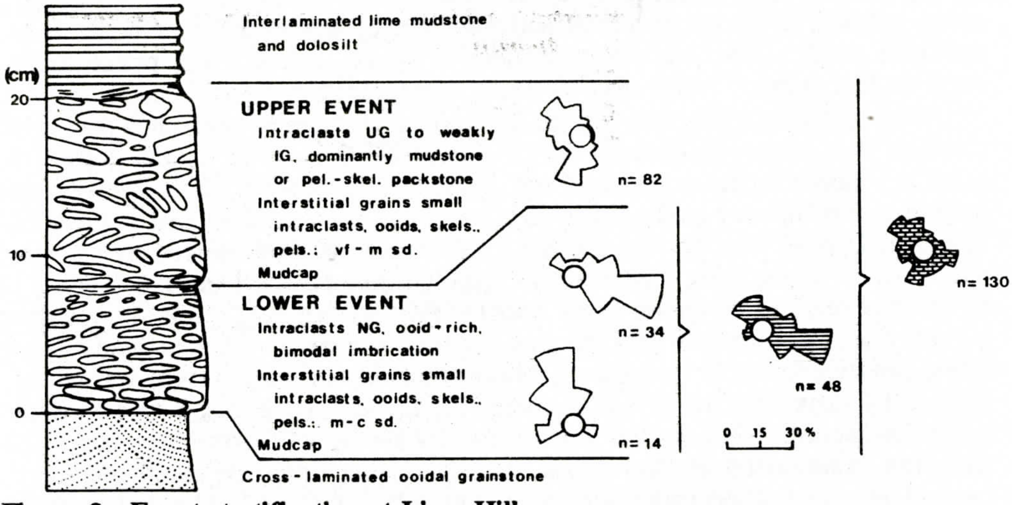

Figure 3. Event stratification at Lime Hill.

The lower event is a grain-supported, normally graded, 8 cm-thick sequence (Figure 3). Intraclast compositions are quite varied and include mixed peloid - bioclast - small intraclast - ooid packstones, ooid grainstones or packstones, structureless mudstone, or bioclast wackestone. The interstitial grains are a poorly sorted mixture of silt-sized peloids and sand-sized ooids, small intraclasts, and bioclasts. The intraclasts and interstitial grains are thus very different in composition and texture from the underlying ooid-rich grainstone. The intraclasts in the lower zone are strongly imbricated and show two distinct orientations (Figure 3). The lower, coarser clasts indicate paleocurrent movement landward whereas the upper, smaller clasts display transport seaward.

The 13 cm-thick upper event suggests similar but somewhat different depositional conditions from the underlying sequence (Figure 3). This conglomerate is grain-supported, ungraded to crudely inversely graded, displays a much greater proportion of mudstone intraclasts, and has finer interstitial material containing a significant mud component. In addition to abundant structureless lime mud intraclasts, some packstone and wackestone types occur also. The interstitial grains are mostly very fine to medium sand-sized and consist of ooids, peloids, small intraclasts, and bioclasts. Patches of mud, much of which is dolomitized, are abundant in the matrix fraction. A thin (one cm) dolomud cap lies at the top of the upper event. Some of the intraclasts protrude into the dolomud cap, suggesting that the high-energy event responsible for transporting the upper conglomerate was followed significantly later by suspension settling of the mud cap without an intervening traction-deposited sand-silt sequence. Above the thin mud cap are several cm of interlaminated lime mud and dolosilt.

Paleocurrent analysis of the upper event shows transport in diverse directions, but most prominently landward and shore-parallel (Figure 3). Note in Figure 3 the paleocurrent information loss when data from various events are combined into one pattern.

The entire sequence described above is interpreted as follows: the basal ooidal grainstone represents normal day-to-day deposition in a well-agitated, subtidal ooid sand shoals environment. The lower conglomerate reflects high-energy storm activity causing rip-up of a variety of subtidal sediments, including ooids, peloids, bioclasts, and mud and transport of this material onto the clean, ooid-rich sand. The bimodal and nearly bipolar paleocurrent pattern again suggests that storm currents may have amplified tidal flood and ebb.

The upper conglomerate formed by storm reworking of substrates somewhat muddier than those represented in the lower event. These might have been subtidal intershoals or shoals mud cap areas, both typically rich in finer-grained sediment. The inverse grading seen in parts of this zone suggests that at times this gravel-rich sheet, like the upper event interval at Eggleston, moved by transport dynamics involving more than ordinary turbulent support mechanisms. Neither of these units seem to be debris flows because they are not very thick, do not show mounded tops nor tapered flow margins (snouts), and have well-organized internal fabric (imbrication) (Hill and others, 1982). Probably the larger clasts were driven upward by dispersive pressures (Bagnold, 1954), suggesting that during transport grain interaction between the intraclasts ("grain flow") was important. (A thorough examination of flow types involved in the origin of conglomerates is beyond the scope of this paper; for such discussions, the interested reader is referred to sources such as Middleton and Hampton, 1976; Lowe, 1982; Harms and others, 1982; and Middleton and Southard, 1984).

Imbrication in the upper Lime Hill event deposit indicates that the direction of movement was generally toward the shore or along the coast. Significantly later after deposition of the coarse material, mud settled onto the intraclast layer which was followed by a return to normal sedimentation of carbonate mud and silt.

Spring City

The last multiple event sequence analyzed is within the Spring City exposure (Figure 1). This section is in one of the most landward Copper Ridge outcrop belts; thus, dolomitization is extensive and quartz sand derived from the craton is abundant.

Three small magnitude events (with possibly a fourth located above the upper zone) are represented within a 20 cm- thick interval (Figure 4). The event sequences are virtually identical and consist of a lower zone of rounded to well-rounded, structureless dolomud intraclasts with interstitial quartz sand that grades upward into quartz arenite. All conglomeratic intervals are grain-supported and grading is normal except in the middle event which is ungraded to crudely inversely graded. The quartz arenites are rippled and contain a few dolomud intraclasts. Desiccation cracks occur in dolomud at the top of the upper event. This dolomud thickens and thins in response to the underlying quartz sand ripple troughs and crests, respectively.

Paleocurrent patterns show a diversity of flow directions within each event (Figure 4). The combined readings suggest net coast-parallel transport.

Each sequence is interpreted as representing intertidal sedimentation during which carbonate mud laminae formed as drapes over underlying rippled siliciclastic sands. The muds commonly dried and cracked and then were

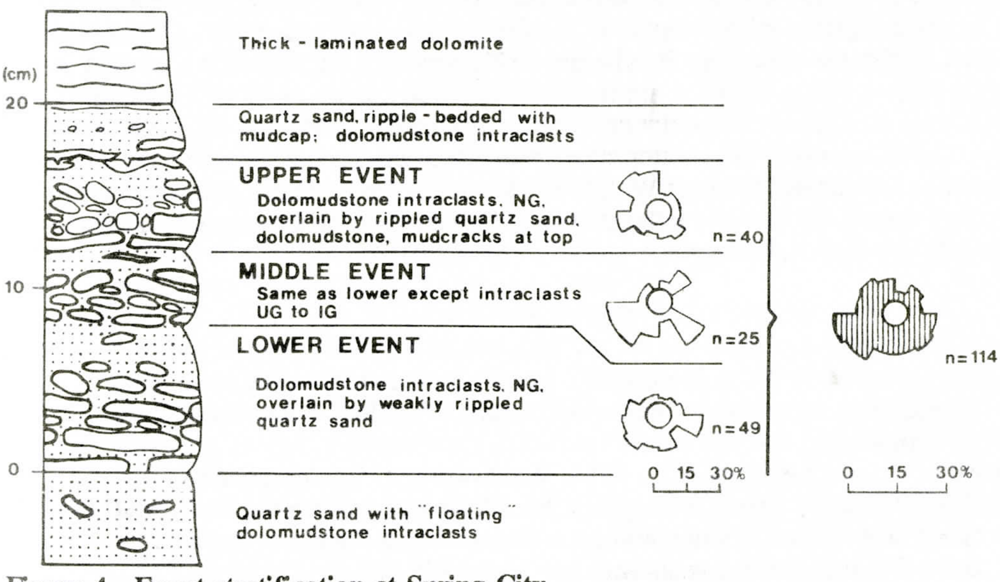

Figure 4. Event stratification at Spring City.

reworked by relatively small-scale storm events that carried the intraclasts in various directions. The ungraded to inversely graded nature of the intraclasts in the middle interval suggests the presence of transport dynamics similar to those inferred for the upper Eggleston and upper Lime Hill event sequences. Quartz sand was mixed with the intraclasts during reworking and rippled quartz arenite caps were deposited by diminishing traction currents. Carbonate mud formed across the siliciclastic intertidal flat to begin the next cycle. Eventually, thick laminated carbonate mud dominated sedimentation as the siliciclastic sand supply was reduced.

CONCLUSIONS

Two important general conclusions are evident from this analysis of event stratification in the Copper Ridge Formation of southwestern Virginia:

(1) Multiple event sequences are common in storm-generated imbricate intraclast conglomerates of the type studied here, and by extension, should be a typical component of peritidal carbonates in general. Indeed, I estimate that about 25 percent of all Copper Ridge intraclast conglomerates observed during this and other studies by me show evidence of multiple event activity.

(2) Important details of genetic interpretation will be lost if multiple event intervals are not individually analyzed. This is especially true when directional data obtained from multiple event sequences are not analyzed as separate events.

ACKNOWLEDGEMENTS

Acknowledgement is made to the Jeffress Memorial Trust (Grant #J-37) and to the Donors of the Petroleum Research Fund (Grant #14785-B2), administered by the American Chemical Society, for their support of this research. I thank my research assistants Jonathan H. Giesen and Gregory L. Carter for their aid in various phases of the work. I am indebted to J. Fred Read, William Koerschner, and Arthur P. Schultz for help in locating the study sections. A constructive review by Kenneth Walker improved the manuscript considerably. I thank Kimbell L. Knight and Stephen W. Lenhart for helpful suggestions about carbonate petrology during the course of the study. All results and interpretations, however, are solely my responsibility.

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